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STORM WARNINGS ON LAKE BALATON,(U)
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FOREIGN TECHNOLOGY DIVISION

STORM WARNINGS ON LAKE BALATON

by

G. Gotz

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STORM WARNINGS ON LAKE BALATON

By: G. Gotz

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STORM WARNINGS ON LAKE BALATON

by G. Götz

Foreword

Lake Balaton is the largest inland lake of Central Europe and is the most important vacation and tourist region of Hungary. Informing vacationers and the various companies concerned with their comfort and entertainment of the expected weather situation, and in addition, the timely warning of bathers and boaters on the lake of impending, dangerous winds, is an important task for our meteorologists. For this reason, we have created an independent storm-warning service within the Hungarian Weather Service, which operates in the summer months in an observatory built in Siófok.

In this volume of the Official Publications of the Central Meteorological Institute, the procedures and prediction methods developed by our collegues are discussed in the interest of developing and improving the activity of the Storm Warning Service. As is known, the prediction of the local weather picture requires knowledge of, and the ability to predict the development and movement of mesosynoptic structures, which are an organic part of the macrosynoptic weather picture. Now this is particularly true for atmospheric processes connected with convective activities which

1. Introduction

T. Tänczer

Due to its geographic location, the area of Lake Balaton assumes an important place among the land features of Hungary. The climate of this region is composed of influences of the flatlands in the East, the gently falling Alpine foreland in the West, and the distant, but still effective action of the Adriatic Sea, where alternately one or the other effect takes precedent. The transitory character of the climate is enhanced by the changing surface structure, by the presence of the Balaton uplands and of the Bakony mountains and these circumstances lead to an even greater multiplicity in meteorologic conditions of the Balaton region. Even the influences of the large quantity of water can be demonstrated in the direct vicinity of the lake, and to be sure, these effects are expressed sometimes in a moderation of temperature extremes and sometimes in the generation of a local breeze. In the daily weather picture, these peculiarities of the climate, resulting from the unique geographic location and deviating considerably from the other land regions, appear in accord with synoptic conditions.

The objective of this publication consists in describing the distinguishing characteristics of the weather conditions of the Balaton region as affected by the lake and its environ. Consequently, the majority of used methods belong to synoptic climatology. With consideration of the fact that practical demands (in the area of recreation, athletics etc.) require a synoptic-climatologic description of the Balaton region primarily for the summer season, the majority of our investigations was dedicated to this period (May-September). In this period, in the interest of recreational safety, a Storm Warning Service is run by the Meteorologic Central Institute. In the writing of this monograph, we endeavored to address the numerous, practical demands. One of the ways leading to the perfection of the Storm Warning Service consists in improving the theoretical level, namely by refinement of objective methods of

storm prediction, or in the preparation of new prediction methods. In this monograph, all research results are presented which may be used in the prediction of storms and of thunderstorm activity at lake Balaton; thus they may serve as a useful aid for employees of the Storm Warning Service.

In the synoptic characterization of Lake Balaton, the emphasis was naturally placed on wind conditions. The wind is the primary principle for sailing on the lake. But the wind is also the main source of danger for the water enthusiasts. Even though the announcement of sailing matches and the establishment of the program for the sailing season only has need of the chronological and spatial distribution of winds, with regard to questions of danger in water sports, the statistics on storms are important. All facts about wind conditions on Lake Balaton are discussed in section 2. Since only the Siófok wind recording instruments are in an outdoor location which permits registration of streaming conditions directly over the lake in the case of a North wind, the named recordings are used as the basis for most data processing used here.

With regard to a definition of storms, a wind threshold value of 15 m/sec is used as a basis. The statistics of storms were produced from observations from the years 1958-1963. Between these years, the distribution of storms is discussed by months, times of day, wind direction, intensity and duration. Sudden storms are distinguished from those with stepwise wind increase. The sudden storms are of obvious importance to water sprots on the lake; these are defined by an increase in wind speed by 8 m/sec in a period of 10 minutes with peak wind velocity reached at 15 m/sec. In such cases, on an initially smooth water surface, breaking waves appear within a few minutes. According to our investigations, the occurrance of this class of storms is linked mainly with thunderstorms. Conversely, the stepwise forming storms often taking hours in formation before reaching storm strength, have no surprise moment and are thus less dangerous.

The storms on Lake Balaton are put into three classes under consideration of their type of formation mechanism: 1. Storms genera-

have a decisive effect on the weather structure during the summer season and which frequently lead to the formation of dangerous storms, devastating cloudbursts and hail. More detailed investigations aimed at an objective prediction of the mososynoptic phenomena were recently begun and at present, the main problem is the recognition of the physical nature of the individual processes and the attempts at a numerical approximation of them. Consequently, the work published in this volume is aimed at creating a foundation for the future preparation of a dynamic model for the mesosynoptic phenomena: The regularities and physical relations are sought which lead to the formation of special wind conditions, to the formation of instability lines. Several statistical studies are included in this volume; these are likewise indespensible for the preparation of a regional prediction.

This work is published in the hope that it will be of interest no only to meteorologists, but also to water enthusiasts.

Budapest, April 1966

Prof. F. Dési, Ph.D.
Director

ted from the instability existing within an air mass, 2. Storms caused by change of air mass, 3. Storms occurring as a result of an increase in air pressure gradients. Among the instability storms are those occurring due to local thunderstorm activity and along the instability lines. Storms with a replacement of air mass are those occurring on passage of a weather front or a frontal zone. The cold fronts moving in from the Northwest undergo a braking in the forezone of the Bakony mountains, then they pounce with renewed strength on the lake surface. The storms generated by an increase in the pressure gradients almost all belong to the group of step-wise developing storms.

In order to perform a spatial exploration of wind conditions on Lake Balaton and to examine the different wind structure at the bank and over the land, the wind data from Siófok is compared with that from Keszthely. The observation post in Keszthely lies at a distance of about 1.5 km from the lake, to the West, on the North side of the Lake. The wind here has a much greater degree of turbulence than in Siófok due to the large undulations of the surface. Consequently, the number of storms is much greater in Keszthely; but this fact may not be attributed to synoptic factors. Another situation exists with regard to the daily distribution of storms, whereby the local differences between the two observation posts come to the foreground. Likewise the consequences of the different orographic conditions are that in Keszthely, besides the northern and southern wind directions, other wind directions are almost non-existant, whereas in Siófok a rather large number of storms come from the western sector.

The wind changes occurring between the banks at Somogyer and Veszprémer were examined by means of measurements taken on-board ships. In 1963 the measurements were taken on the ferry route of Siófok-Balatonfüred and the results are also reported here. It is notable that in the case of a North wind, the values of wind velocity in Balatonfüred are an average of only 60% of those in Siófok.

Attempts to predict wind conditions on Lake Balaton are presented in section 3. In the method used, the air-pressure difference of Szonbathely-Kecskemét is related to the northern winds in Siófok. In another method, the values of the air pressure gradients occurring in a circle of 150 km around Siófok which exceed the amount of 2 mb/300 km, are considered and their relation with the maximum wind speed occurring within three hours is examined. From this investigation result the intensity differences between landward and lakeward winds and the differences in the deviations of the geostrophic winds which are generated as a result of the differently developing friction forces. An investigation of the isallobar wind component led to a reinforcement of the Ertel description over that of Brunt and Douglas, according to which a decrease in wind is caused by a coincidence of the baric and isallobaric gradients, but an opposing direction of the two vectors is supposed to result in an increase in wind. The Ertel description is based essentially on the fact that the wind needs a certain adaptation time compared to the pressure fields. For the size of this adaptation time, we obtained a value of about one hour.

Besides the prediction of the macroscopic wind, we also performed investigations on the prediction of wind blasts coming from non-frontal thunderstorms. The used method is based on experiences of Brancato that a close relation exists between the max. wind velocity due to a convective thunderstorm on the one hand, and with the temperature decrease occurring with the wind blast on the other hand. An acceptable result on the temperature decrease was obtained by means of the moist-potential temperature in the condensation level.

To predict the stepwise developing storms, the method of Litvinova is used; it is based on the change in pressure field using the tendency picture. By testing this method with regard to the winds in the Balaton region, its utility in storm warnings was enhanced.

Since the sudden outbreak of storms is usually connected with a thunderstorm, a broad area was made available for the treatment of the question of thunderstorms on Lake Balaton (section 4). If we

examine a thunderstorm map of Hungary, then it is striking that the Balaton region has a preferred position there; this is explained primarily by the orographic location of the Balaton region. A refined distribution of thunderstorms on Lake Balaton also seems to indicate this. One special investigation is concerned with the synoptic circumstances of the outbreak of thunderstorm, i.e. with questions of synoptic objects connected with the particular thunderstorm.

In section 5 we are concerned with the possibilities of thunderstorm prediction. Previously, the best results were obtained by the objective (statistic) prediction methods. So it was obvious that we should also perform tests with such a method. Special attention was paid to the changes in the status curve; these occur within the prediction period. Trials with the best-known instability indices were performed. The best results were obtained with the index of Showalter, provided its geostrophic advection is also considered. In contrast to the original criterion, we obtained the result that under present climatic conditions, instead of the value 0, a value of +2 announces the appearance of a thunderstorm. We succeeded in improving the results obtained by use of the Showalter index by including another element in the method, i.e. a so-called multi-parameter alternative method was used.

As is known, the instability does not always lead to the initiation of thunderstorm activity. We also performed tests aimed at exploring the reasons for this phenomenon and found that in the majority of cases, the lack of moisture is responsible for this state of affairs. Another paper is concerned with the influence of the structure of the wind field and wind shear on the structure of thunderstorm activity.

As was mentioned, the most powerful and dangerous storms on Lake Balaton originate by the passage of instability lines. Consequently, a detailed and large-scale investigation concerning the origin of the instability lines and their passage across Hungary--especially over the Balaton region--was performed (section 6).

It turned out that the vast majority of instability lines passing over the transdanube region occurs in a special synoptic situation, namely when a depression is over the Po valley and at the same time, an anticyclonal wedge is penetrating in from West Europe over Austria with a cold front on its leading edge. Here, in the Carpathian basin there is a warm, moist subtropical air mass having a particularly pronounced tendency to form thunderstorms. At the end of that section, the method used in the USA to predict instability lines is presented and it provides the potential for drawing conclusions about the formation of an instability line on the basis of the 12-hour changes in the absolute topography 850 mb and under consideration of the humidity conditions.

Due to the water quantity of Lake Balaton, the temperature conditions of the direct vicinity are affected. This effect is especially pronounced on sunny days when the local lake wind system can be fully unfolded. Investigations on the daily course of water and air temperature and on the question of the interaction of air and water were performed in section 7.

In the present monograph, it was our intention to touch on all important questions of the synoptics of Lake Balaton, but this work cannot be considered exhaustive. With regard to the wind conditions of Lake Balaton, it would be necessary to perform detailed aerial photography. Under this circumstance and having only extensive observation material, we can make a well-founded guess. It would be desirable to perform simultaneous wind measurements in the environ of the lake and on the lake itself. Similarly for the daily temperatures, since they are likewise very dependent on local conditions. Further progress may be expected in the area of practical prediction methods for the individual weather elements. Through individual mesosynoptic investigations, the problems can doubtless be further clarified. These problems have to do with the generation and course of the various storms.

2. Wind Conditions of Lake Balaton

2.1 Introduction (B. Bójti and G. Götz)

If a ranking of meteorologic elements is set up in the investigation of weather at Lake Balaton, then no doubt one arrives at the conclusion that the most important element is the quasi-horizontal air movement, i.e. the wind. This prominent role of wind conditions is due primarily to the effects of the wind on water sports and on the lake itself. The nearly 600 km^2 of lake surface is in direct contact with the atmosphere and any change in the latter causes large or small changes in the water mass of the lake. Under the action of the wind, the shoals have breakers and the water level increases, in comparison to the horizontal equilibrium state at that bank of the lake exposed to a landward wind. Observed differences in water level in a cross-direction between the Balatonakali and Balatonszemes observation posts, were move than 40 cm under the effect of a North wind, whereas for air motions in the longitudinal direction, differences of 55-65 cm were measured.

The wind causes waves on the water surface; a characteristic shoal wave pattern forms: The waves are short and steep and they quickly form breakers. The wave height depends on the length of the water segment exposed to the action of the wind. Due to fluctuation in wind strength, different forms of wave fronts occur. But it is very interesting that the widely accepted view in laymens circles of 1 meter high waves on Lake Balaton, is greatly exceeded by measurements: On July 8, 1963 at Balatonszemes, in a North wind gust at a speed of 28 m|sec, a wave height of 1.8 m was measured by the recording instruments of the Research Institute for Water Resources (VITUKI). The breakers on Lake Balaton are primarily dangerous to sailboats since they are easily tipped. The decline in water motions caused by the wind is not simultaneous with the decline of the wind, rather there is a certain time shift.

The wind investigations have a practical significance and a meteorological importance. In accord with the objectives of the work, the majority of the effort by the Storm Warning Service on

Lake Balaton is aimed at dissemination of information and prediction of wind conditions. The daily work of the Warning Service is composed of the following tasks: Several daily wind forecasts, issuance of alarm warnings announcing the impending arrival of storms in the lake region, information and advice to the water Police, for ship travel, for sailing associations, for other Organs and to private individuals; in all these activities, a detailed knowledge of wind conditions on Lake Balaton is indespensible. Namely, through the special orographic conditions, an investigation of the properties of the air streams appearing over the lake is required.

The first specialized wind investigation performed with a view toward storm warning, comes from Ambrózy, Götz and Tänczer [1]. In that paper, as an initial step in the area of investigation of wind conditions, the properties of the sudden storms were examined on the basis of data collected in the course of five summer months from the years 1957-1961 in Siófok. Through this work, the principle and practical foundations for a synoptic classification of sudden storms was created; the chronology, distribution by wind directions and velocities and the most important synoptic circumstances of the sudden storms were examined in detail.

The first detailed wind statistics collected with a view toward storm warnings, comes from Graics and Kálmán and is based on observations made in the course of five summer months from 1958-1962 [2]. The goal of this work was to obtain an overview of the monthly frequency, chronciogy, wind direction and velocity conditions of storms on Lake Balaton. As the basis of the data processing, the wind recordings of the Keszthely and Siófok observation posts were used, and thus it became one of the primary tasks of the work to explain the differences in wind conditions at the two observation posts.

After these two basic works, several investigations followed; they were designed to describe and define the special streaming conditions in the Balaton area. These investigations created the indespensible conditions for the advisory service of the Storm Warning Service, and also the foundation for research work performed

in connection with wind prediction.

2.2 Statistic and Synoptic Investigations of Storms on Lake Balaton (G. Götz and T. Tänczer)

In the present paper, the properties of storms over Lake Balaton are examined on the basis of observations of the five summer months (1 May to 30 Sept.) from the six years 1958-1963. The use of a longer time period was not possible since sometimes no observations were available, and sometimes no reliable data was available. The processing of this data is based on the recordings of the individual observation posts who recorded the actual wind conditions on Lake Balaton, namely, the wind recordings of the Fuess wind instruments at the Observatorium Siófok. In the study period, one other wind recording instrument was in operation in Balatonkenese, but the results cannot be used due to an instrument error. A representative, operable wind recording instrument was put in operation in 1963 in Balaton-szemes by the Research Institute for Water Management (VITUKI), whose data will in future be a useful supplement to our observations. The data of the wind instrument of the meteorological observation post in Keszthely can only be used conditionally for Lake Balaton; a comparison of wind conditions of Siófok and Keszthely is presented in section 2.3. and 2.4.

The present investigation is organized into four parts. Section 2.2.1 contains a statistic handling of all storms recorded by the instrument of the Observatorium Siófok, with respect to their chronology, duration, direction and intensity. In section 2.2.2 the storms are classified on the basis of their outbreak, in sec. 2.2.3 on the basis of their synoptic character. In section 2.2.4, some phonemena accompanying storms are investigated.

2.2.1 Statistic Properties of the Storms

In this paper, a Storm is a wind period during which the max. instantaneous wind gusts reach a threshold value of 15.0 m/sec or more. We distinguish between the beginning of the wind period (outbreak of wind) and the beginning of the storm. The beginning of the wind period is defined by that timepoint when the effect of the synoptic system begins, by which the wind is generated. This

timepoint can in most cases be determined objectively, at least it is distinguished by a general wind rotation and by a slow or rapid increase in wind velocity. The outbreak of the storm is defined by the first wind gust to reach the value of 15 m/sec, and the end of the storm by the last such wind gust. The duration between these two times is called the storm duration. An independent wind period, or an independent storm, is that wind caused by a given synoptic system. However, within one and the same wind period, two or more storm periods are viewed as independent if two wind gusts having a speed of 15 m/sec or more, are separated by a time span of more than 6 hours. Accordingly, two sequential storms are separated either by a wind rotation or by a storm-free period lasting at least 6 hours. The latter case is relatively rare and occurs only for a special type of storms (in a situation with a so-called Azores advance).

Table 1: Distribution and Number of Storms in Siófok

	1 Mai	2 Juni	3 Juli	4 Aug.	5 Sept.	Σ
1958	6	7	7	7	1	28
1959	4	11	10	4	1	30
1960	7	12	11	14	4	48
1961	8	8	10	6	3	35
1962	10	10	8	8	6	42
1963	10	4	3	9	7	33
Σ	45	52	[illegible]	48	22	216
6 Mittel	7,5	8,7	8,2	8,0	[illegible]	36,0

Key: 1-May 2-June 3-July 4-August 5-September 6-average

Using the above definitions, we obtained a figure of 216 independent storms for the studied period. The annual and monthly distribution is presented in table 1. It is notable that in 1960 there was an abundance of storms with a very large number in the three summer months. From a summation of results it follows that on average for the month of June, the greatest number of storms occurs, and the smallest number is in September. This is explained by the large frequency of monsoonal thunderstorms connected with storms in June and by the frequent, anticyclonal weather situation in September ("Old wives summer").

Table 2: Duration of Storms in Hours and in Percent for the Observation Period in Siófok.

	1 Mai	2 Juni	3 Juli	August	September	Σ
1958	34,4 4,6%	94,6 13,1%	36,5 4,9%	74,4 10,0%	2,2 0,3%	242,1 6,6%
1959	22,9 3,1%	114,6 15,9%	47,2 6,3%	72,9 9,8%	25,0 3,5%	282,6 7,7%
1960	59,1 7,9%	80,7 11,2%	150,1 20,2%	45,0 6,1%	7,7 1,1%	342,6 9,3%
1961	54,4 7,3%	32,5 4,5%	31,5 4,2%	45,3 6,1%	11,0 1,5%	174,7 4,8%
1962	61,9 8,3%	94,3 13,1%	5,7 0,8%	33,2 4,5%	16,2 2,2%	211,3 5,8%
1963	39,1 5,3%	6,3 0,9%	21,6 2,9%	46,6 6,3%	34,6 4,8%	148,2 4,0%
Σ	271,8 6,1%	423,0 9,8%	292,6 6,6%	317,4 7,1%	96,7 2,2%	1401,5 6,4%

Key: 1-May 2-June 3-July

An important supplemental quantity for the number of storms is the duration of each storm. In this regard, table 2 provides some information in which the duration of storms is expressed in percent of the total length of the particular time period. In this regard too, the year 1960 proved to be the stormiest year of the period: The total duration of storms was 342.6 hours, or 9.3% of the total time there was a stormy wind. In July of that year, the total duration of storms was 150.1 hours corresponding to 20.2% of the total time of the month, whereas in Sept. 1958, the total duration of time with 15 m/s wind was only 2.2 hours, and this corresponds to only 0.3% of the total time of the month. In the summation, the months of June and September have the extreme values and the average duration of storms is 6.4% of the total time. The average duration of a storm was 6.5 hours.

The picture of the wind conditions over Lake Balaton can be refined if we examine how the number of storms is distributed by duration (table 3). Accordingly, half of the storms last less than 2 hours, three-quarters last less then 8 hours, and 90% of them do not last more than 20 hours. Only 2% last more than 40 hours,

Table 3: Frequency Distribution of Storms by Their Duration in Siófok

1 Zeitdauer	1 Min.	2—15 Min.	15—30 Min.	0,5—1St. 3	1—2	2—3	3—4	4—6	6—8	8—10	10—15	15—20	20—25	25—30	30—40	>40 St. 3
2 Zahl der Stürme	26	26	15	24	15	14	13	14	13	12	16	7	7	4	5	5
%	12	12	7	11	7	7	6	7	6	6	7	3	3	2	2	2

Key: 1-duration 2-number of storms 3-hours

and one-third of storms last only 1 to 15 minutes. There is a large number of storms which last only one minute (with the instantaneous wind velocity exceeding a value of 15 m/s only once or twice). The longest storm occurred during the study period of July 1960; this storm lasted from 23 July 0434 hours until 26 July 0543 hours (Central European Time), or 73.2 hours of stormy wind. A storm of similar magnitude occurred in 1964, but it does not belong to the study period. In this case, the storm broke loose on 21 Sept. at 0557 hours and lasted until 24 Sept. 0445 hours: The stormy North wind lasted 70.8 hours and several times exceeded a value of 20 m/s and on 23 Sept. at 1145 hours, it reached a value of 27 m/s. Within the study period there were two other long storms on 16-18 August 1959 and on 11-13 June 1958, with a duration of 49.9 hours.

Table 4 provides information about the frequency distribution of the maximum wind velocities occurring during the storms. In 68% of storms, the wind speed remained below 20 m/s and only in 2% did the value exceed 30 m/s. A storm of this intensity can also occur in September (25 Sept. 1963), and wind gusts above 35 m/s occurred in the course of the study period only in the months of July and August.

Within an independent wind period, the wind direction does not change to any notable extent if we are dealing with a longer wind period. The direction of the storm can be best recognized by the direction of the max. wind gusts, and this identification remains

Table 4: Frequency Distribution of Maximum Wind Speeds of Storms in Siófok.

Windgeschwindigkeit 1 (m/sec)	2 Mai	3 Juni	4 Juli	Aug.	Sept.	Σ Stürme 5	Σ %
15,0—19,9	35	32	30	33	16	146	68
20,0—24,9	8	13	15	9	5	50	23
25,0—29,9	2	6	2	5	·	15	7
30,0—34,9	·	1	1	·	1	3	1
≥35,0	·	·	1	1	·	2	1

Key: 1-wind speed 2-May 3-June 4-July 5-storms

valid--with a few exceptions--for the duration of the storm. For this reason, in the investigation of storm directions, the direction of the maximum wind velocity is used (table 5). The most striking property of this problem consists in the fact that more than three-quarters of all storms (78%) have a direction from the sector NW-N, whereby during the study period, not one storm arrived from the direction ESE-SSE. A more thorough interpretation of this table, and of the secondary maximum in WSW can only be given on the basis of the synoptic classification of storms; this will be done in section 2.2.3.

Table 5: Distribution of Storms According to the Directions of Maximum Wind Gusts in Siófok.

Richtung 1	Mai 2	Juni 3	Juli 4	Aug.	Sept.	Σ Stürme 5	Σ %
S	1	·	·	·	·	1	0,5
SSW	1	2	1	3	·	7	3
SW	4	1	·	3	1	9	4
WSW	·	2	5	5	1	13	6
W	1	2	1	·	·	4	2
WNW	1	3	4	2	·	10	4
NW	3	6	9	6	3	27	12,5
NNW	12	17	14	15	4	62	29
N	19	17	15	14	12	77	36
NNE	1	·	·	·	1	2	1
NE	·	1	·	·	·	1	0,5
ENE	·	1	·	·	·	1	0,5
E	2	·	·	·	·	2	1
ESE	·	·	·	·	·	·	·
SE	·	·	·	·	·	·	·
SSE	·	·	·	·	·	·	·

Key: 1-direction 2-May 3-June 4-July 5-storms

Table 6: Distribution of Maximum Wind gusts According to the Wind Direction in Siófok.

Richtung [1]	S	SSW	SW	WSW	W	WNW	NW	NNW	N	NNE	NE	ENE	E	ESE	SE	SSE
15—20 m/s	1	6	5	8	2	6	19	39	55	2	1	0	1	.	.	.
20—25 m/s	0	0	3	2	2	4	6	16	15	0	0	1	1	.	.	.
25—30 m/s	0	1	0	0	0	0	2	6	6	0	0	0	0	.	.	.
30—35 m/s	0	0	0	1	0	0	0	1	1	0	0	0	0	.	.	.
≥35 m/s	0	0	0	2	0	0	0	0	0	0	0	0	0	.	.	.

Key: 1-direction

Also, the distribution of max. wind velocities according to the various wind directions was examined; this is shown in table 6. It is notable that--although the greatest part of the weak, medium-strength and strong storms have a direction from the NW-N sector--the strongest storms come from the WSW direction. The explanation of this circumstance is also given in section 2.2.3.

For the work of the storm warning service, for the dissemination of prognostic information and for the solution to numerous practical problems, the following factors have a great significance: The timing of storm outbreak, the time of strongest wind and in general, the distribution of the storms to the individual hours of the day. But these questions cannot be answered in a concrete and practical, useful way without a causal classification of storms (except for a few general findings. Storms can originate at any time of day and the maximum wind speed can be attained any any hour of the day. When broken down hour-by-hour, it was found that most storm outbreaks occur in the period between 19 and 20 hours or between 22 and 23 hours (each 7% of the total storms), whereas the smallest number of outbreaks occurs from 12 to 13 hours (2%). If three-hour time segments are used, then most storms break out between 18 and 21 hours (17%), with the fewest between 10 and 13 hours (7%). For a six-hour break-down, we obtain extreme values for storm beginnings between 17 and 23 hours (33%) or between 7 and 13 hours (17%) respectively. The least reliable findings are obtained by comparing halves of a day, since then we find that more than half of all storms, or 61% break out in the time from 13 to 01 hours.

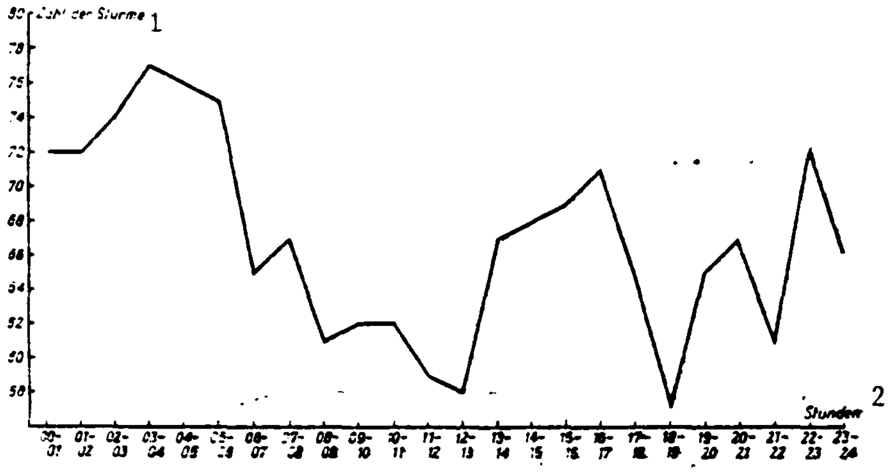


Fig. 1: Number of Storms in Siófok Occurring in the Individual Hours of the Day for the Study Period 1958 to 1963.
Key: 1-number of storms 2-hours

For a one-hour organization of storms, the max. speed is in most cases (but here too, only 7% of all storms) in the time between 4 and 5 hours, or in the time between 19 and 20 hours, whereas between 12 and 13 hours, only 1% of storms have their max. speed. If we use 3-hour segments, then the maximum is between 14 and 17 hours (18%), for 6-hour segments, the max. is between 14 and 20 hours (32%), whereas in the time from 10 to 13 hours, only 6% have their maximum, and between 7 and 13 hours, only 17% have a maximum. If a half-day split is used, then for storm intensity we get the same result as for their outbreak: 61% of storms attain their greatest intensity in the time from 13 to 01 hours.

In the investigation of storm frequency in the individual hours of the day, likewise no significant differences are found. Relatively smaller is the storm appearance from 18 to 19 hours and from 12 to 13 hours; in these two hours storms occur in only 6.2 or 6.3% of all investigated cases. The hour with greatest storm occurrance is from 3 to 4 hours with a frequency of 8.4%. In supplement to this data, it should be noted that storms make up an average of 7.3% of all time. Most storms occur between 00 and 06 hours, the smallest number occur in the time from 11 to 13 hours. Storm occurrance in the individual hours of the day is shown in fig. 1.

2.2.2 Differentiation of Storms by Type of Outbreak

Among those storms which reach the lake, the most dangerous ones are those which break out suddenly, i.e. have properties that shortly after outbreak of the wind, it reaches its maximum speed. According to the definition of the World Meteorologic Organization, an air stream is called a sudden outbreak of wind when the wind speed increases from its outbreak to at least 8 m/s in a period of 10 minutes, whereas a sudden storm occurs if the wind speed becomes 15 m/s. In the present investigation, this definition is also used as the basis for the classification of storms, but it should be noted that according to experiences of the Storm Warning Service, much smaller increases in wind can in many cases be dangerous for boats of all types, especially sailboats.

Table 7: Distribution of Storms According to Type of Outbreak in Siófok

	1 Mai	2 Juni	3 Juli	Aug.	Sept.	Σ
P	13	14	16	18	2	63
P in Prozenten von Σ 4	29	27	33	38	9	29%
S	32	38	33	30	20	153
S in Prozenten von Σ 4	71	73	67	62	91	71%

Key: 1-May 2-June 3-July 4-in percent of

Table 8: Distribution of Wind Velocity Values for Sudden Storms (P) in Siófok

Windgeschwindigkeit 4 (m/sec)	1 Mai		2 Juni		3 Juli		Aug.		Sept.			
	P	5 in Prozenten von Σ	P	5 in Prozenten von Σ	P	5 in Prozenten von Σ	P	5 in Prozenten von Σ	P	5 in Prozenten von Σ	P	5 in Prozenten von Σ
15,0—19,9	9	28	9	29	9	30	9	27	0	0	36	25
20,0—24,9	3	39	3	23	5	33	5	56	2	40	18	36
25,0—29,9	1	50	2	33	0	0	3	60	·	—	6	40
30,0—34,9	·	—	0	0	1	100	·	—	0	0	1	33
≥35,0	·	—	·	—	1	100	1	100	·	—	2	100

Key: 1-wind velocity 2-May 3-June 4-July 5-in percent of

If storms occurring within the study period are broken into two groups of sudden storms (P) and stepwise forming storms (S), then we obtain the classification shown in table 7. Thus, not quite one-third of all storms belong to the class of sudden storms, and they occur in greatest numbers in July and August, and least in Sept. If the

distribution of max. wind velocities for sudden storms is examined (table 8), then we see the following: The larger the speed interval taken, the more storms P become prominent over storms S. Whereas storms with max. velocity less than 20 m/s were only one-third of sudden outbreak, among the storms with max. velocity from 25 to 30 m/s, even 40% were of sudden outbreak; and speeds over 35 m/s occur exclusively in suddenly occurring storms. Thus, the sudden storms are usually stronger than developing storms. The reasons for this must lie in the differences between the atmospheric processes causing the two types of storms; this is discussed in section 2.2.3.

The direction distribution of sudden storms is given in table 9. Sudden storms can come up from all directions. In fact, during the study period, all storms coming from S, W, NE and ENE were exclusively sudden storms. Conversely, although most storms P have a NNW or N direction, 18% of all storms come from those directions. Conversely, it is striking that the storms P from the WSW have a large frequency. This direction distribution can likewise only be explained on the basis of the storm generation mechanism.

Table 9. Distribution of Sudden Storms (P) by Wind Directions in Siófok.

Richtung 1	S	SSW	SW	WSW	W	WNW	NW	NNW	N	NNE	NE	ENE	E	ESE	SE	SSE
P	1	2	5	11	4	5	5	11	14	1	1	1	1	.	.	.
In Prozenten 2 der Σ der Stürme	100	40	56	85	100	50	19	18	18	50	100	100	50	—	—	—
In Prozenten von P 3	2	3	8	17	6	8	8	17	22	2	2	2	2	—	—	—

Key: 1-direction 2-in percent of all storms 3-in percent of P

2.2.3 Classification of Storms by Their Underlying Synoptic Mechanisms

Through a suitable, statistic handling of the data obtained in the course of the Universal wind measurements, a detailed and true picture of the particular wind conditions of a given region can be outlined. But in order to have a better explanation for the most important lines of this picture and to have a deeper insight into

the differing physical circumstances of the individual storms, the storms must be classified by type of triggering, atmospheric mechanisms.

A wind of storm strength can be generated in the air layers near the ground, through various atmospheric processes which can be precisely defined. These processes (and the storms connected with them) can be classified in various manners depending on the goals of the particular investigation. The principles of the classification contain the fact that the storms generated from identical or similar atmospheric processes, should be ordered in the same class so they can be compared in this manner with storms of other classes.

Under consideration of the goals of our investigations, the storms penetrating into the area of Lake Balaton can be organized into three main classes, the one main class and two subclasses. In our classification we wanted to be sure that every storm could be classified and that the classification be based on the synoptic situation and behavior of meteorologic elements in an objective manner. In this classification, main classes were formed from the following, basically different-structured storms: Frontal storms with an air-mass replacement; baric storms caused by the convection activity. Thus, there results the following outline for the classification of storms:

Storms with replacement of air mass		L
1) Frontal storms	Lf	
2) Frontal zone storms	Lz	
Baric Storms		B
1) Prefrontal storms	Bp	
2) Gradient storms	Bg	
Instability storms		I
1) convective storms	Ik	
2) Squall line storms (instab. lines)	Is	

In the course of storms L, the air mass initially present is replaced by a different air mass having a different thermodynamic and hydrodynamic structure (in general, the second air mass is

Table 10: Frequency of Storm Types

Art des[1] Sturmes		2 Mai		3 Juni		4 Juli		August		September		Σ	
		Stürme 5	%	Stürme 5	%	Stürme 5	%	Stürme 5	%	Stürme 5	%	Stürme 5	%
L	Lf	12	27	12	23	13	27	21	44	9	41	67	31
	Lz	12	27	17	33	7	14	10	21	8	36	54	25
	Σ	24	54	29	56	20	41	31	65	17	77	121	56
B	Bp	4	8	2	4	·	—	4	9	1	5	11	5
	Bg	8	18	12	23	16	32	4	9	3	13	43	20
	Σ	12	26	14	27	16	32	8	16	4	18	54	25
I	Ik	9	20	7	13	9	19	3	6	·	—	28	13
	Is	·	—	2	4	4	8	6	13	1	5	13	6
	Σ	9	20	9	17	13	27	9	19	1	5	41	19

Key: 1-type of storm 2-May 3-June 4-July 5-storms

colder). The replacement of air mass can be accompanied by a sudden change in state, whereby the interface between the two air masses is small and the new air mass proceeds on a sharply-defined front. In this case, jump-like changes in the quantities of state are noted; the wind velocity generally rises quickly (storm Lf), but the storm wind is usually not of long duration. However, the air-mass replacement can also proceed slowly; in this case, the two air masses are separated by a broad frontal zone. The characteristic values for the new air masses in this case are taken only stepwise from the quantities of state, likewise the wind slowly attains storm intensity (storm Lz), but the storm often lasts longer than storm Lf.

Baric storms are generated in the macroscopic air-pressure fields within one and the same air mass, due to the increase in the air-pressure gradient. Since the increase in baric gradient is a slow process, the winds B have the property to undergo a stepwise increase in strength. Storms B can be generated within the warm sector of a cyclone, in the area in front of the cold front (storm Bp) or even in regions free of fronts (storm Bg).

Storms I are likewise winds generated within one and the same air mass. The most freqeunt variant is storm Ik which occurs in connection with thunderstorm activity. The processes occurring within the cumulonimbus lead to an accelerating, descending air stream which is forced to spread out when it reaches the ground leading to a fast

increase in the wind. Since this spreading air mass can move out up to 80 to 100 km from the base of the thundercloud, it is possible that a storm Ik can occur in those regions where the observation points have observed no thunderstorm. The most dangerous and at the same time rarest type of storms are storms Is. They are generated by a special weather situation, in front of a powerful cold front, within the warm sector, due to the development of a linearly arranged convection activity (see section 6). The outbreak of this storm is sudden in all cases, often without the appearance of any sign, and the most powerful storms occurring around here are generated in this manner.

In table 10, the frequencies of individual storm types are given. According to the table, the wind in most cases is a result of the replacement of air mass and within this class, the frontal effects usually cause the storm intensity. More than half of all storms occur at the fronts and frontal surfaces; one fourth of them result from baric influences and only the remaining 19% of storms occur from instability storms. Rarest are squall line storms (6%) and the prefrontal storms (5%). Essentially we find the same distribution within the individual months. The number of storms L decreases toward high summer; in Sept., 77% of all storms are caused by frontal influences. Conversely, the summer is the season for instability storms. In September, there are no more convective storms and a squall line was observed only once in September during the entire investigation period. The fraction of baric storms is nearly the same in each month.

In table 11, the frequency distribution is shown for the value intervals of the maximum wind velocities for the various storms. The most powerful storms with winds exceeding 35 m/s were caused without exception by squall line processes. This is the only type of storm where the max. frequency is not also the lowest velocity step: 85% of the squall line occurrences in the area of Siófok caused a storm of at least 20 m|s. The weakest storms are generated from prefrontal processes, usually they do not even attain a velocity of 20 m/s. The increase in baric gradient and the simple convective activity do not lead to wind gusts exceeding 30 m/s. Within the class of

Table 11. Distribution of Maximum Speeds of Various Storm Types. (Z= number of storms, HG=frequency percentage of velocity, HS= frequency percentage of storm type).

Geschwindigkeit 1 (m/sec)	L: Lf Z	L: Lf HG %	L: Lf HS %	L: Lz Z	L: Lz HG %	L: Lz HS %	L: Σ Z	L: Σ HG %	L: Σ HS %
15,0—19,9	39	27	58	38	26	70	77	53	64
20,0—24,9	19	38	28	12	24	22	31	62	25
25,0—29,9	8	53	12	3	20	6	11	73	9
30,0—34,9	1	33	2	1	33	2	2	67	2
≥35,0	.	.	.	.	.	.	.	.	.

Geschwindigkeit 1 (m/sec)	B: Bp Z	B: Bp HG %	B: Bp HS %	B: Bg Z	B: Bg HG %	B: Bg HS %	B: Σ Z	B: Σ HG %	B: Σ HS %	I: Ib Z	I: Ib HG %	I: Ib HS %	I: Is Z	I: Is HG %	I: Is HS %	I: Σ Z	I: Σ HG %	I: Σ HS %
15,0—19,9	9	6	82	33	23	77	42	29	78	25	17	89	2	1	15	27	18	66
20,0—24,9	2	4	18	9	18	21	11	22	20	2	4	7	6	12	47	8	16	20
25,0—29,9	.	.	.	1	7	2	1	7	2	1	7	4	2	13	15	3	20	7
30,0—34,9	.	.	.	.	.	.	.	.	.	.	.	.	1	33	8	1	33	2
≥35,0	.	.	.	.	.	.	.	.	.	.	.	.	2	100	15	2	100	5

Key: 1 - velocity

storms L, the frontal influences are in general, linked with stronger winds than dictated by the influences of frontal zones, though both processes have caused storms with winds of more than 30 m/s.

Table 12: Wind-Direction Frequencies of Various Storm Types

1 Richtung	L				B				I			
	Lf		Lz		Bp		Bg		Ik		Is	
	Z	HS%	Z	HS%	Z	HS%	Z	HS%	Z	HS%	Z	HS%
S	.		.		.		.		1	4	.	
SSW	.		.		4	36	.		2	7	1	8
SW	.		.		4	36	.		2	7	3	23
WSW	.		.		3	28	.		5	18	5	38
W	.		.		.		.		2	7	2	15
WNW	2	3	1	2	.		2	5	4	14	1	8
NW	6	9	6	11	.		12	28	2	7	1	8
NNW	28	42	19	35	.		11	25	4	14	.	
N	29	44	27	50	.		17	40	4	14	.	
NNE	.		1	2	.		.		1	4	.	
NE	.		.		.		.		1	4	.	
ENE	1	1	.		.		.		.		.	
E	1	1	.		.		1	2	.		.	
ESE	.		.		.		.		.		.	
SE	.		.		.		.		.		.	
SSE	.		.		.				.		.	

Key: 1-direction

A differentiation of wind directions by type of storm (table 12) offers a much more valuable picture than table 5. The most important and striking circumstance is that the various types of storms are groups around fixed directions. The vast majority of storms L (85-86%) come from direction N or NNW, and storms Bg generally come from the same direction. Now since these types of storms account for 76% of all storms, this direction sector also is prominent in table 5.

The role of the NW-N sector for the types of storms in question is linked with the typical weather situation of the Carpathian basin. The fronts of cyclones moving in the broad, western flow, reach the trans-Danube plain generally from a NW or NNW direction, and after passage of the cyclone, a "backside situation" results which causes the typical structure of the baric N-winds. An unusually strong increase in the air-pressure gradient occurs as a rule, in such a manner that to the West of Lake Balaton, in the regions of the Alps and in Austria, a powerful pressure increase occurs, whereas in the

East, the pressure is unchanged, or decreases (weather situation with an Azores air-pressure wedge). This isallobar influence also leads to North winds. Naturally, synoptic situations occur which differ from this typical, regular situation. In past years, in the second half of May, regular, so-called East weather situations formed. The storms coming from these time periods appear in the columns ENE and E of the table. These cases form striking properties of the table, and even the synoptic situation which formed them is unusual.

The direction SSW-WSW of the prefrontal storms is likewise connected with the typical structure of the synoptic situation. At the front side of the West-to-East moving, powerful cyclones (North of our latitudes) and in the warm sector of these cyclones, the wind is turning stepwise from SE to SW. The air pressure gradient reaches a value in the second half of this period, which is needed to form a storm. At this time, the wind direction is still SW. Since in prefrontal weather situations the baric gradient rarely reaches the value needed for a storm, prefrontal storms lying outside the sector SSW-WSW, simply do not occur.

The directions of the squall-line storms move between broader limits, but essentially around the direction WSW. In those special cases in which a squall line is formed within the Carpathian basin (described in detail in sec. 6), the passage of the line of convective activity over the land is usually in a SW-NE direction and leads to maximum wind gusts from WSW and SW. In this manner, the mentioned fact that the most powerful storms come from the WSW direction, takes place (table 6).

The only type of storm having no pronounced directional preference is the storm Ik. Convective storms can come up from all directions, except from the sector ENE-SSE. This relatively uniform distribution is a result of the fact that convective thunderstorms can form at any direction from the observation post and consequently the direction of the storm generated by this thunderstorm, can be quite variable. The reasons for this must lie in the geographic conditions as described in the referenced paper [1]. The storm-free

sector corresponds to the environ of Siófok which lies inland since winds generated there experience a strong braking effect through friction on the ground. This explanation is supported by the fact that the convective storms coming from the sector S-SW, which also lies in the direction inland, have wind velocities which in no case reach 20 m/sec.

Table 13 gives an overview of the duration of storms of various type. It is striking that although the frontal storms have a shorter duration in accord with their properties, than the frontal zone storms, nonetheless, cases of frontal storms of very long duration do occur; even those storms having the longest duration in the course of the observation period, belong to the frontal storm group. This fact indicates that the method of storm typing used here could be improved in this regard. Namely, in the creation of the classification of storms, as mentioned, we proceeded exclusively from the processes by which storms are generated, thus only those synoptic circumstances are used which are responsible for the outbreak of the storm. But now there are mixed types. For example, frequently the outbreak of a storm occurs under frontal effects, but then under the influence of a post-frontal, baric action, it takes on a very long duration. This case occurs often when an easterly propagation meets a westerly anticyclone. The cold air mass in the stream field of the anticyclone, frequently meets a sharply defined front, and in the wedge-shape forming behind the front, very long-lasting storms can form (even those storms having the longest duration) due to the gradients. In such cases, the storm (in accord with circumstances for its outbreak) can be called a frontal storm, even though many hours later the continuing storm no longer has any relation with the front. Naturally, such a classification of storms can be structured which will take this circumstance into account; see e.g. sec. 2.4.

Table 13. Duration of Various Storms

Dauer 1	*Lf*	*Lz*	*Bp*	*Bg*	*Ik*	*Ie*
Mittel 2	[illegible],3 St. 5	9,1 St. 5	3,7 St. 5	7,0 St. 5	13 Min.	21 Min.
Minimum 3	1 Min.	1 Min.	1 Min.	1 Min.	1 Min.	1 Min.
Maximum 4	73,2 St.	49,9 St.	9,9 St.	38,6 St.	49 Min.	1,4 St.

Key: 1-duration 2-average 3-minimum 4-maximum 5-hours

Long-lasting storms can also be caused by baric effects which are not related to a change of air mass. Among these, the gradient storms has a longer duration, whereas the duration of a prefrontal storm is usually determined by the appearance the cold front.

The instability storms have a short duration. For the convective storms, the max. wind gust in many cases is only one to two times the value of 15 m/sec and the whole storm lasts only 1 to 2 minutes. This type of storm with an average duration of only 13 minutes forms the shortest group of storms, and not one case has been found in which such a storm lasted one hour. Also, the squall line does not last long. A squall line storm of one hour is a rarity. The case of 13 July 1961 with a storm lasting 1.4 hours was the result of an exceptionally broad zone of convective activity.

Table 14. Classification of Various Storm Types by Type of Outbreak

	Lf		*Ls*		*Bp*		*Bg*		*Ik*		*Is*	
	Z	%	Z	%	Z	%	Z	%	Z	%	Z	%
P	22	33	·	0	·	0	1	2	27	96	13	100
S	45	67	54	100	11	100	42	98	1	4	·	0

In the evaluation of the data in table 14, in which the various types of storms are classified by their type of outbreak, the strictness of the criterion by which a "sudden" storm is defined, must be pointed out. Thus, we have the state of affairs that even fronts linked with a rapidly unfolding change in air mass, can only be called "sudden" in one-third of cases. On the other hand, it is very remarkable that in the course of the study period, a Bg storm occurred which is a "sudden" storm. This case occurred on 3 July 1959, when the influence of the baric gradient was suppressed by a countereffect generated from a mesosynoptic disturbance. In the wedge situation of a SW anticyclone, like that prevailaing over the entire Carpathian basin with a lively NW wind corresponding to the macrosynoptic situation, in the morning hours there formed a South to Southwest flow in the southwest parts of the Transdanube basin. In the course of the late morning, the wedge situation had progressed so much that at 1100 Central European Time (CET) the air pressure gradient reached a value of 1.5 mb/100 km over the land, which gave

a NW gradient wind of storm strength. The force of this powerful, macrosynoptic effect was soon felt. The wind jump occurred in the environ of Siófok at 1123 CET at 1127 hours, the NW storm broke loose and at 1202 hours, a wind gust of 18.8 m/sec was noted from this direction.

The instability storms have almost without exception, a sudden outbreak and the squall line storms are all "sudden" storms. The reasons for this must be sought in the outbreak mechanism of Storm I.

Through the investigation of the outbreak of the various storms and of the time of max. wind gust, one arrives at the conclusion that the storms Lf, Lz and Bg begin at any time of day and can attain maximum strength at any time. We have more detailed information only for storm Lf. From the viewpoint of these storms, the time between 11 and 13 hours seems to be calmest: During the observation period, in the time from 11 to 13 hours, no Lf storm broke out and not one Lf storm attained its max. strength during this time. Of the frontal storms, a relative maximum of outbreaks (25%) occurs between 03 and 06 hours, but there is a large number of frontal storms (23%) which begin between 18 and 21 hours. During the night, between 18 and 06 hours, the outbreak of more than 3/4 (76%) of frontal storms occurs. Their max. strength (Lf storms) is generally reached in the time between 22 and 01 hours (25% of storms), whereas between 11 and 14 hours, the greatest development of these storms was noted in only 4% of cases. With regard to storms Lz and Bg, not as much information is available. This blurring of the daily profile is a result of the nature of these storms: The fact that their outbreak and profile are dependent on macrosynoptic conditions; but these latter are subject to such influences which largely are unrelated to the time of day. Storms Bp all break out between 08 and 14 hours and the max. strength is always reached between 09 and 15 hours. This could be explained by the fact that at the time of beginning of intense solar radiation, the thermal opposition existing in the region of the cyclone undergoes an intensification, thus also increasing the baric gradients and the wind can more

easily reach storm strength. Proof of this is found in the weather situations in which the prefrontal effect continues for a long time. In such cases, the wind direction SW continues for several days, but its velocity reaches storm strength only in the daytime; at night there is a decrease. A weather situation at the beginning of summer 1964 provides an excellent example of this (this period is not in the observation term). The prefrontal effect began in Siófok on 1 June at 0800 CET. There was a stepwise increase in the SW wind and at 15 hours, wind gusts of 15 m/s were measured. Then, a decrease occurred, but on 2 June, a new increase in wind velocity was noted after 03 hours; at 10 hours, a wind of storm strength was blowing and at 15 hours, a speed of nearly 20 m/s was reached. Then, a decline in the wind occurred again, but in the daytime hours of 3 June, there was a storm again. The 66 hour-long prefrontal effect was stopped by the arrival of a cold front on 4 June at 0203 CET.

The storms Ik break out in a time beginning at 13 hours and 89% of them break loose during this period which extends out to 23 hours. The remaining, small number of storms Ik broke out in the midnight hours and in the early hours before daybreak. In the time between 04 and 13 hours, not one convective thunderstorm broke out during the investigated period. In the majority of convective thunderstorms, the beginning is between 16 and 18 hours, and the same finding also applies to the timing of maximum storm intensity. The explanation of this distribution is due to the fact that the max. of atmospheric instability occurs during this time and this state often leads to the formation of a convective thunderstorm, which is then linked with thunderstorm wind gusts.

The greatest number of storms Is break out between 14 and 21 hours (69%), the other storms Is begin in the hours around daybreak. With the exception of squall line storm of 19 August 1960 (which broke out at 1020 CET), no storms Is began in the time between 05 and 14 hours. Due to the shortness of storms Is, the same findings apply to the timing of max. storm strength. As we see, we have no such prominent picture as in the case of storms Ik, since Is storms

are affected by the instability of the air column and the mesosynoptic factors, which depend only slightly on the solar radiation.

Table 15. Characteristics of Extremely Powerful Storms in Siófok

1 Datum	2 Beginn des Windes	3 Beginn des Sturmes	4 Zeitp. des max. Windstosses	5 Max. Windstoss		8 Sturmart	9 Dauer des Sturmes
				6 Richtung	7 Geschwindigkeit (m/sec)		
19. Aug. 1960	10·20	10·20	10·24	WSW	36,8	Is	9 Min.
13. Juli 1961	14·10	14·10	14·15	WSW	36,2	Is	84 Min.
11. Juni 1959	10. Juni, 03·50	10. Juni, 08·14	06·00	N	31,8	Ls	47 St.
22. Juli 1958	18·55	18·56	19·00	WSW	31,0	Is	21 Min.
25. Sept. 1963	19·50	20·34	23·27	NNW	30,9	Lf	17 St.

Key: 1-date 2-beginning of wind 3-beginning of storm 4-time of max. wind gusts 5-max. wind gust 6-direction 7-speed 8-storm type 9-duration of storm

Finally, in table 15 we find the most important data for the five most powerful storms occurring in the course of the observation period. From this comparison it follows that the conditions and prerequisites for the outbreak of a 30 m/s storm are quite different.

2.2.4 Some Phenomena Accompanying Storms

In the paper [1] the question of the internal cohesion is discussed in detail as it pertains to the sudden onset of storms and thunderstorm activity. The mechanics and thermodynamic processes unfolding inside the thundercloud, form practically the sole factor which can lead to the sudden outbreak of a storm. This fact was reinforced by the present investigation. Within the study period there was a simultaneous thunderstorm activity within a circle of 100 km of Siófok, i.e. in a space in which the wind of a thunderstorm head can exert an influence according to the investigations (except for the storm Bg of 3 July 1959 which was discussed above). In the course of these sudden storms, in 60% of cases a thunderstorm was observed from the observatory itself, in another 18%, thunderstorms occurred in a circle of 10 km. In 10% of cases, a thunderstorm occurred in a circle of 20 km, in 6% in a circle of 40 km. In the remaining 6% of cases, the thunderstorm was more than 40 km away, but it was still within a circle of 100 km. From this fact it is

quite clear how important it is to know all important properties of the thunderstorm and to take into account the improving methods of thunderstorm prediction used by the Storm Warning Service for Lake Balaton.

The question of what values of air pressure (converted to sea level) give a storm outbreak was examined. The result runs as follows: In the studied period, the outbreak of storms occurred for air pressure values between 993 and 1026 mb. The most frequent air pressure value for the time of storm outbreak is 1012 mb (in 12% of cases), whereas nearly half (48%) of storms lie within the interval of 1010-1015 mb. It was further found that 79% of storms belong to air pressure values between 1008 and 1019 mb, whereas at a pressure above 1020 mb, only 4% of storms began. If the air pressure has a value above 1026 mb, then no storm occurs. The storms occurring at high pressure (above 1021 mb) were type Bg winds; thus we find that in the course of an increase in the baric gradient, even at high pressure, a storm can occur. At very low air pressures, storms occur which are linked to an isallobar field with quickly falling air pressure. Such storms are: Frontal storms, prefrontal storms and squall line storms.

2.3 Statistics of Summer Storms around Lake Balaton (É. Kálmám and Á. Graics)

In this section, investigations of a statistical nature are presented; these are based on wind recordings of the Keszthely and Siófok observation posts. Information processing was performed on data from the periods of 1 May-30 Sept. for the five years of 1958-1962. The objective consisted primarily in obtaining an overview of the differences in wind conditions at Keszthely and Siófok.

In a comparison of wind conditions at Keszthely and Siófok, the difficulty arises that the set-up of wind instruments at the two observation posts is very different. In Siófok the instrument is completely free; it protrudes up over the lake, but in Keszthely, the instrument is located 1.5 km from the bank, in a built-up area, on a smaller hill, and thus serves rather to reflect wind conditions of a city region. Even though the buildings nearby are not nearly as

high as the instrument, it is possible that the surrounding buildings and the orography could generate a powerful turbulence. For the city set-up of the wind instrument in Keszthely, it is observed that the recording pen itself moves back to zero position after the strongest wind gusts. However, in the wind recordings in Siófok, the conditions in the free air flow over the lake are reflected; for landward winds, the pen can stay above the zero line for a long time.

With regard to the different equipment set-up and because of the interests of the Storm Warning Service, in the evaluation we did not use the average wind speed, rather we used the envelope curves of the wind gust recordings. In the evaluation, each wind increase received special attention provided the max. wind gust at one or both observation posts was measured as more than 15 m/s, and also when it lasted even only one minute. In addition, every wind increase was considered to be an independent storm when the average value of wind gusts reached a value of 13-14 m/s for at least 1½ hours. Accordingly, as the time of outbreak, we did not use that time when the first 15 m/s wind gust occurred, but rather that time when the average value of 13-14 m/s began. This time was determined with an accuracy of 10 minutes.

This selection of threshold value is based on the pertinent regulations of the Storm Warning Service, which specify that a pre-alarm rocket will be fired for a wind gust of 12 m/s, and at 15 m/s a full-storm alert must be called.

The duration of storms defined in this manner was likewise determined with an accuracy of 10 minutes. In cases where only one single wind gust occurred, the storm duration is assumed to be one minute. In such cases where the wind velocity declined for 1 to 2 hours below the threshold value of 13-14 m/s, and then rose again, these two, or possibly three, waves were not separated, but rather treated as a single storm. As wind direction, we used that direction prevalent at the beginning of wind increase.

In the course of the 25 investigated months, in Keszthely 186

storms and in Siófok 182 storms were registered by the wind recording instrument. There were 61 storms in Keszthely alone, and 57 storms in Siófok alone. According to this data, it can be maintained that the storms in Siófok and in Keszthely occur in nearly equal numbers. If we consider the entire region of Lake Balaton, then on average ten storms per month are expected.

The frequency curves of storms are presented in fig. 2. To make this figure, the storms were split into four groups according to their spatial size. In group 1 are storms occurring either in Siófok or in Keszthely, or at both places simultaneously; these are the cases in which there was a storm somewhere on Lake Balaton. In group 2 are storms limited to the northern basin of Lake Balaton represented by Siófok, and group 3 consists of storms occurring only in the southern basin of the lake, represented by Keszthely. Finally, group 4 contains those storms registered simultaneously at both observation posts, and thus extended simultaneously over the whole Balaton region.

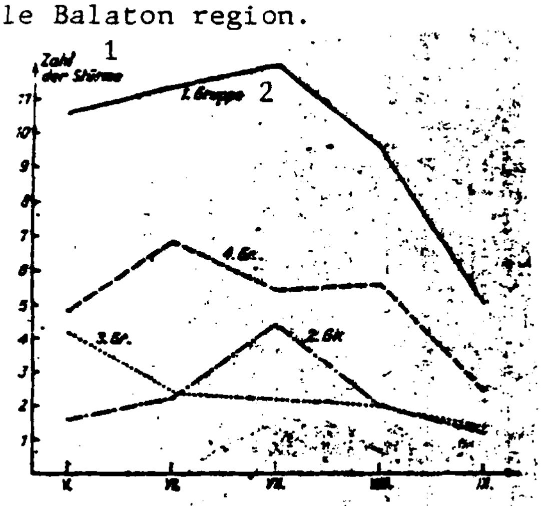


Fig. 2. Frequency of Storms on Lake Balaton

Key: 1-no. of storms 2-group

From fig. 2 we see that in general, the month of July has the greatest number of storms. The average number of all storms in this month is 12, whereas in May and June, it is 11, in August, 10, in September, only 5. The number of storms extending over the entire Balaton Lake region is about half as large as the total number of storms. The spatial extent of storms in the individual months is represented by the deviations between curve 1 and curve 4. In May, June and July, there are large deviations between the two curves

and wind increases of a local nature, connected with convective thunderstorms, are very numerous in those months. In the course of August and September, the two curves come closer together. In these months, the number of storms is smaller, but they usually extend to the entire Balaton region.

For a comparison of wind conditions in the northeast and southwestern basin, curves 3 and 2 are suitable; these denote storms occurring only in Keszthely or only in Siófok. On average, in each month, 2-3 such storms are expected to extend only to one part of Lake Balaton. The increase in the number of storms occurring only in Siófok in July is notable; in this month on average four such storms occur which are observed in the northeast basin of Siófok or perhaps only in the environ of Siófok. This phenomenon can probably be explained by local factors.

The classification of storms by their duration is an important problem. To persons relaxing in the Balaton region or to sport sailors it is an important matter to know if we are dealing with one or two strong wind gusts, or with a long-lasting wind. On the basis of such thinking, the storms are broken into four groups according to their duration:

1. Storms lasting about one minute; this group contains storms lasting 1-2 minutes. From a synoptic standpoint, these represent two types of storms for which the short duration is a characteristic feature: a) A wind gust suddenly developing from a weak wind and usually caused by a local, convection thunderstorm; b) A wind gust developing from a lively wind, which exceeds 15 m/s and then continues for a time.

2. Storms lasting about 10 minutes. In this group are storms lasting 10 minutes to max. one hour. These can be air-mass storms or frontal storms.

3. Storms lasting about an hour. This group is defined so that it contains storms lasting from 1 to 6 hours.

4. Storms lasting about one day. This groups consists of storms

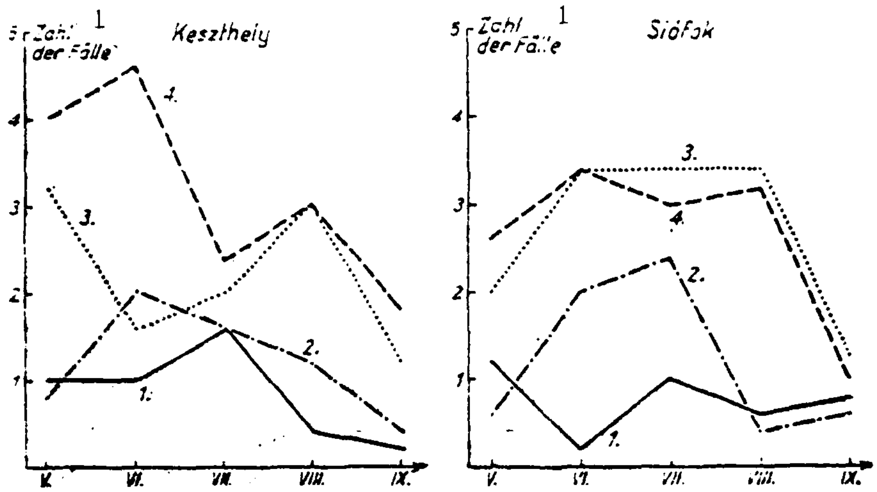


Fig. 3. Frequency of Storms according to their Duration: 1. Minutes; 2. 10 minutes; 3. Hours; 4. Days

Key: 1-number of cases

lasting more than 6 hours. As mentioned, long-lasting storms are considered to be a single storm if for one to two hours a decline in the wind led to wind speeds below the threshold value, but then the threshold value was reached again.

From fig. 3 we see the monthly frequencies of storms of different length. The most striking property of this series of curves consists in the fact that the storm frequency increases with storm duration at both observation posts. The wind gusts lasting one minute have the lowest frequency. On average, only one such storm is expected each month. The wind increases lasting 10 minutes are more frequent. In July and August, two such storms are expected each month. The number of storms lasting about an hour or a day is nearly 3 on a monthly average from May to August, but this figure drops to 1 in September. The storms occurring in Sept. are generally of short duration. But these late-summer thunderstorms also have the property that they extend over the entire Balaton region. The number of storms lasting half a day or a whole day are most frequent.

Long-lasting storms occur in Keszthely smoewhat more frequently than in Siófok. Their average number in Keszthely is 3.2, and in Siófok, 2.6. For instance, in Keszthely there was a storm lasting

from 29 May 1960 at 0140 CET for 63 hours (more than two and one-half days) with wind gusts of 20 m/s, whereas in Siófok only on 31 May at 07 hours CET was an average wind gust of 14 m/s measured for a period of half an hour.

In Keszthely, the longest-lasting storm observed in the course of the observations began on 1 June 1962 and lasted 87 hours with a max. wind speed of 25 m/s. In Siófok, the longest-lasting storm occurred in 1960 and began in the early morning hours of 23 July and lasted 80 hours. In this storm, the max. wind gust reached a value of 27 m/s.

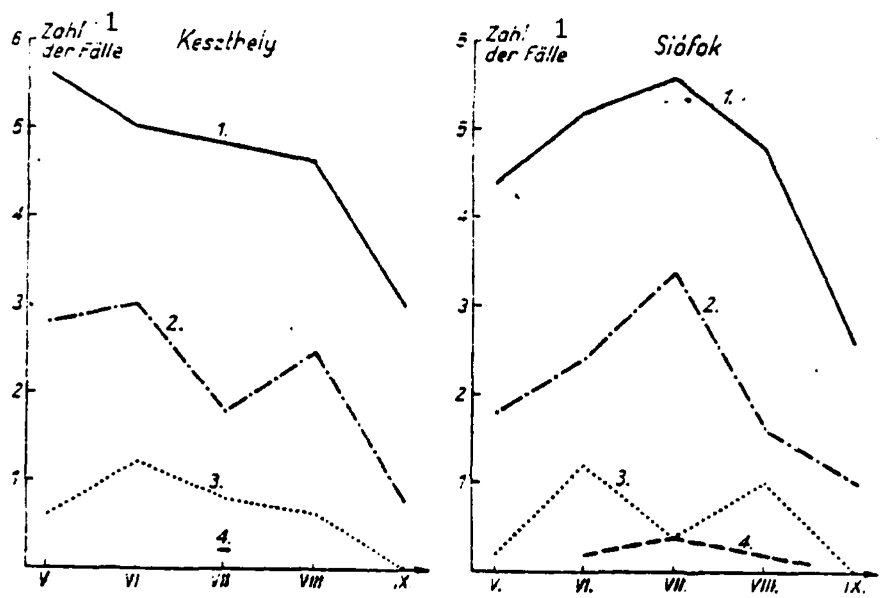


Fig. 4. Storm Frequency According to Maximum Wind Gusts: 1. 15-19 m/s; 2. 20-24 m/s; 3. 25-29 m/s; 4. more than 30 m/s

Key: 1-number of cases

In order to examine the frequency distribution of storms by their intensity, the storms were split into four groups on the basis of the size of the occurring, maximum wind gust. The first group contains those cases where the max. wind gust lies between the threshold value and 19 m/s; the second group contains those between 20 and 24 m/s, group three has 25 - 29 m/s storms, and group four is composed of storms with the speed of max. wind gust exceeding

30 m/s. The frequency distribution of max. wind gust for the individual groups is shown in fig. 4. This figure cannot provide much characteristic curves. Of course, the storms having only smaller wind gusts are most frequent, and with an increase in wind strength, there occurs a gradual decrease in storm frequency. The monthly average storm frequency in which the max. wind gust reaches 19 m/s, is 5, i.e. in each summer month in the area of Lake Balaton, five storms of this intensity are expected. According to the monthly frequency curves, in September, the smallest number of such storms occurs. The average monthly frequency of wind gusts of 20 - 24 m/s is 2 to 3, whereas storms of 25 - 29 m/s average only one each month. Finally, a 30 m/s or more wind gust in Keszthely was noted only once in 5 years: on 13 July 1961 with a maximum wind value of 35.8 m/s. Conversely, in Siófok, there were four such cases. Thus, in Siófok every summer, a 30 m|s or more, wind gust is expected. The max. wind gust observed in Siófok was 36.8 m/s and it occurred on 19 August 1960.

In the course of the 5 years of study, there were 61 such storms over Lake Balaton which were recorded only in Keszthely. The average of max. wind gusts of these storms is 17 m/s, thus they are not particularly strong. The number of those storms occurring in the area of Siófok was 57 for the 5 years. The average, max. wind gusts in this case was 17 m/s. The storm of 5 July 1958 should be mentioned: Here the max. wind gust in Siófok reached a value of 22 m/s, whereas in Keszthely at the same time, the instrument recorded a wind calm. In addition, there were 3 other cases during the 5 years when Siófok has a storm and Keszthely had a calm at the same time.

The distribution of storms by time of day (outbreak) is shown in fig. 5. In Keszthely, the outbreak of most storms is between 7 and 8 o'clock. In the other hours, we see a notable decrease in storm frequency, then there is a secondary maximum at 19 hours. The picture for Siófok is less clear; there is a more uniform distribution of storms to the various times of day. Four maxima are noted: at 8-9, 13-16, 19 and 22 hours.

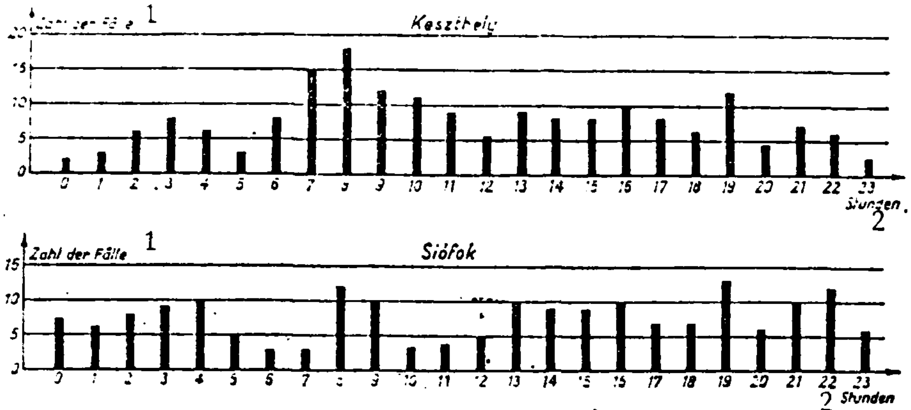


Fig. 5. Hourly Frequency of Storms

Key: 1-number of cases 2-hours

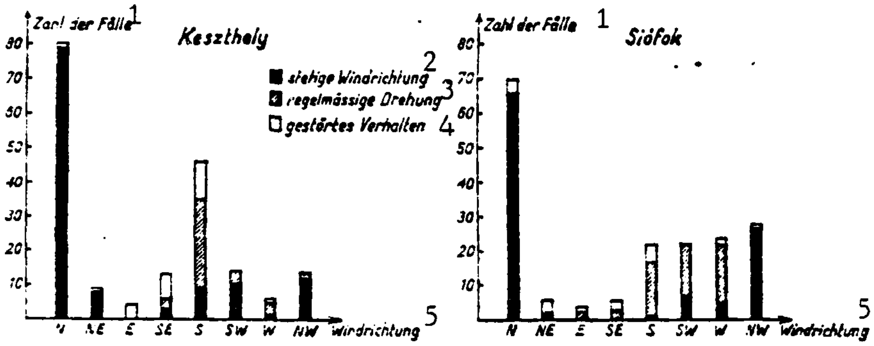


Fig. 6. Distribution of Storms by Wind Direction

Key: 1-number of cases 2-constant wind direction 3-regular rotation 4-discontinuous behavior 5-wind direction

The distribution of storms by wind direction is shown in fig. 6. As mentioned, we used that wind direction as the basis of these calculations, which was observed on outbreak of the particular storms. The cases in which the wind continued from a single direction, i.e. no rotation or significant fluctuations occurred, were assigned to group 1. Group 2 includes those storms where a wind rotation characteristic of front passage occurred. In such cases, we used the wind direction prevailing before the storm threshold was crossed. Finally, group 3 holds storms having irregular, repeated fluctuations in wind direction on outbreak. A general conclusion valid for both observation posts is that storms with a Northerly direction are the most frequent. Another preferred wind direction

is the South. The largest part of southern storms arrive with a regular wind rotation. The number of storms beginning with an irregular wind rotation is negligible compared to the number of the other two groups. This type of discontinuous behavior of wind directions occurs mainly for storms from a southern and eastern direction. The smallest number of wind gusts comes from the East. Most of the irregular changes in wind direction occur during the initial period of the storm.

In a comparison of the number of northern and southern storms we find that the ratio of these numbers shows a shift such that in Keszthely, the southern and in Siófok the northern storms are somewhat more frequent. In Keszthely, all Northern storms have this direction. In Siófok we see that a small number of northern storms undergoes a direction change. Keszthely contains a very small number of Westerly storms; in Siófok the number of such storms is four times as great, and the affected storms usually have a regular wind rotation.

Finally, we tried to answer the question of the direction and speed of storms in the area of Lake Balaton. The time differences between the outbreak of individual storms was studied in Keszthely and Siófok. If there was a time difference in outbreak of more than 6 hours at the two observation posts, then these cases were treated as if they were two different storms.

According to the results, the number of wind increases of West-East moving storms is on average twice (75) that of East-West moving storms (38). The interval between the two posts is traversed in both directions in an average 1½ hours, but the scattering about this average is very high in individual cases--regardless of time of year. Storms coming from the East have their greatest frequency in June.

2.4 Some Comparisons Between Wind Conditions at Siófok and Kesz .ely (G. Götz)

To develop the differences between the wind conditions of Siófok

and Keszthely, another type of processing was performed, based on observations from the period of 1 May to 30 Sept., 1963. This investigation differed from the work reported in section 2.3. Its task was to analyze finer details of the streaming properties at the two observation posts which are considered representative for the two basins of Lake Balaton, and to investigate the dependence of deviations on the macrosynoptic situation which causes the wind. We consider the use of the method applied here, primarily as an attempt only, since the investigation period is much too short to be able to derive any final conclusions.

As a basis of the data processing, a detailed analysis of the recordings of the Fuess wind recording instrument in Siófok and Keszthely was used. The investigation was performed with regard to Strong Winds and Storms. Strong winds were defined as gusts of 12.0 m/s, and storms were wind periods during which the max. wind gust reached a value of-15.0 m/s.

Table 16. Number of Strong Winds and Storms in Siófok and Keszthely (1963)

Geschwindigkeit (m/sec) 1	2 Mai	Juni 3	Juli 4	Aug.	Sept.	Σ
			Siófok			
≥12,0	17	20	5	15	9	66
≥15,0	10	4	3	9	7	33
			Keszthely			
≥12.0	19	22	11	23	9	84
≥15,0	12	12	6	12	4	46

Key: 1-speed (m/s) 2-May 3-June 4-July

The number of strong winds and storms at the two observation posts is shown in table 16. According to the table, the environ of Keszthely has a windier character than the area of Siófok, and this can be explained from the known fact that in Keszthely, the winds are gustier due to the instrument's location and storm strength of individual wind gusts occur for a smaller, average wind speed. The question of gusts will be discussed below and in section 2.5.

The wind conditions of a location are not entirely defined by the number of strong winds and storms. The duration of a certain wind speed is a very important criterion, and this can be determined

with great accuracy from the Fuess recordings. The results obtained for the two observation posts are presented in table 17. The more windy character of Keszthely is reflected also by this data and by the duration of the winds which exceed the threshold values, and also by the average duration of strong winds and storms. However, the very strong, 25 m/s storms in Siófok have a greater frequency, which is explained by the orographic properties of the region.

Table 17: Duration of Winds Exceeding Various Velocity Limits, 1963.

1 Geschwindigkeit (m/sec)	2 Mai	3 Juni	4 Juli	Aug.	Sept.	Σ
			Siófok			
≥12,0	76,9 St.	20,9 St.	26,4 St.	61,2 St.	59,8 St.	245,2 St.
≥15,0	31,4 St.	1,0 St.	17,6 St.	19,2 St.	15,8 St.	85,0 St.
≥20,0	5,3 St.	2 Min.	1 Min.	1,4 St.	6,5 St.	13,3 St.
≥25,0	2 Min.	.	.	.	2,7 St.	2,7 St.
≥30,0	.	.	.	.	3 Min.	3 Min.
5 Durchschnittliche Dauer des starken Windes	4,5 St.	1,0 St.	5,3 St.	4,1 St.	6,6 St.	3,7 St.
6 Durchschnittliche Dauer des Sturmes	3,1 St.	15 Min.	5,9 St.	2,1 St.	2,3 St.	2,6 St.
			Keszthely			
≥12,0	127,1 St.	12,3 St.	73,2 St.	79,7 St.	32,9 St.	324,9 St.
≥15,0	76,7 St.	50 Min.	45,0 St.	41,2 St.	15,7 St.	179,4 St.
≥20,0	5,6 St.	.	15,2 St.	6,0 St.	4,5 St.	31,3 St.
≥25,0	1 Min.	.	7 Min.	.	.	8 Min.
≥30,0	.	.	.	.	.	.
5 Durchschnittliche Dauer des starken Windes	5,7 St.	34 Min.	6,6 St.	3,5 St.	3,7 St.	3,9 St.
6 Durchschnittliche Dauer des Sturmes	5,4 St.	4 Min.	7,5 St.	3,4 St.	3,9 St.	3,9 St.

Key: 1-velocity 2-May 3-June 4-July 5-average duration of strong wind 6-average duration of the storm. St.= hours

Table 18. Percentage Distribution of Strong Winds and Storms by Wind Directions, 1963

Richtung 1	S	SSW	SW	WSW	W	WNW	NW	NNW	N	NNE	NE	ENE	E	ESE	SE	SSE
Siófok	3	2	9	3	2	2	7	18	32	3	2	5	5	3	2	.
Keszthely	4	1	5	4	2	1	5	12	49	4	.	.	10	2	.	1

Key: 1-direction

The frequency distribution of strong winds and storms according to direction of the maximum wind gust is presented in table 18, where the North direction in more prominent in Keszthely than in Siófok:

Nearly half of all maximum wind gusts came from this direction. Due to the shielding effect of the Keszthely mountains, not one strong wind gust came from the directions NE and ENE, but the East winds have a great frequency--this direction corresponds to the Bank line of Keszthely--Balatongyörök. Regarding distribution of wind directions in Siófok, refer to section 2.2, and for Keszthely, refer to section 2.3.

At the beginning of storms, an increase in wind gusts from 5 m/s to 10 m/s is reached in Siófok within an hour on average, but in Keszthely within 3/4 hour on average. For an increase of wind gusts from 10 m/s to 15 m/s, a time of 2½ hours is needed in Siófok and in Keszthely, two hours is needed.

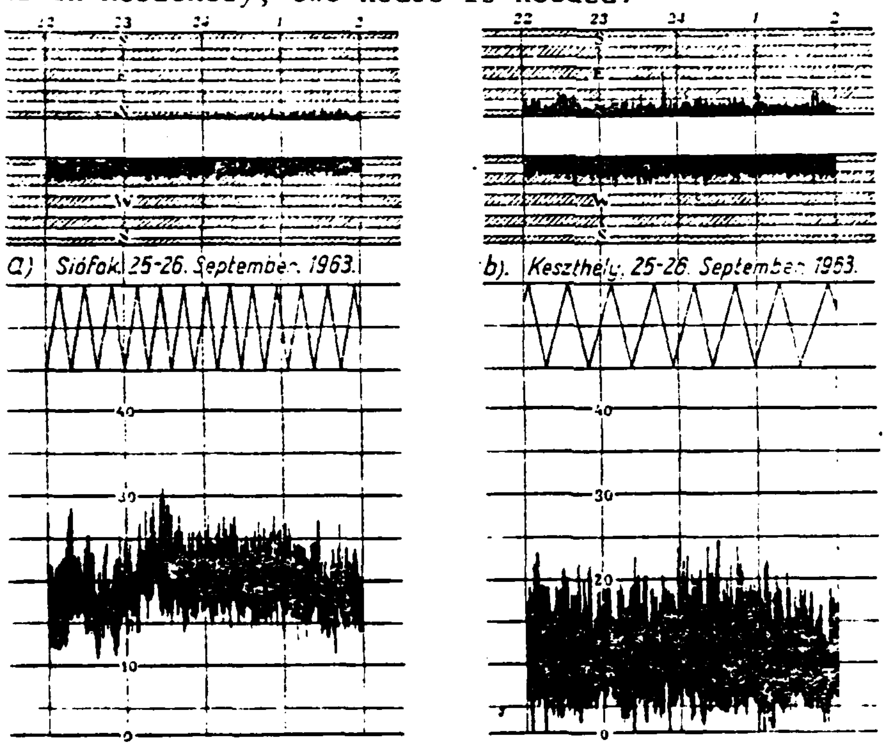


Fig. 7: Structures of Stormy Wind at Keszthely and Siófok during the Storm of 25-26 September, 1963

At the end of the storms, a decline in wind gusts from 15 m/s to 10 m/s occurs on average in 1.7 hours in Siófok; in 2.0 hours in Keszthely. In the course of the storms an average 25 degree fluctuation of wind direction was noted at the two observatories. The fluc-

tuation of wind speed in the course of storms is more characteristic: In Siófok, it was 6.5 m/s and in Keszthely, 12.4 m/s, and this gives another characteristic value for the essentially greater gustiness of the wind in Keszthely (fig. 7). The velocity fluctuations in Siófok can also be considerably high in the case of land winds, but storms from those directions are rare, as presented in section 2.2.

For a deeper understanding of the properties of storms and to interpret the results, it is important to differentiate storms according to the synoptic situation, as was done in sec. 2.2. In that section, a classification of storms based on their origin was performed and some deficiencies of this classification were pointed out. In the course of this work, another system of classification was suggested, which is presented in table 19. This new system consists of a greater number of classes which permits storm differentiation from the viewpoint of closely related synoptic situations, but which have essentially different wind structures.

Table 19. Storm Types

I.	Frontal winds	F	(Lf)
II.	Frontal-baric winds	FB	
	1. Anticyclonal wedge situation	FBa	(Lf)
	2. Backside location in cyclone	FBc	(Lf)
III.	Baric winds	B	
	1. Prefrontal winds	Bp	(Bp)
	2. Anticyclonal wedge situation	Ba	(Lz)
	3. Backside location in cyclone	Bc	(Lz)
	4. Gradient increase for other reasons	Bg	(Bg)
IV.	Convection winds	K	(Ik)
V.	Squall line winds	S	(Is)

The frequencies and some properties of the various storm types were summarized in tables 20 and 21 for the area of Siófok or Keszthely. Strong winds occur at both observation posts most frequently in the train of convective activity, where powerful wind gusts are generated primarily in the direct vicinity of the cloud mass due to the spread of cooled air falling to the ground from the thundercloud. These wind increases occur almost exclusively--as section 2.2 shows--in a sudden manner and consequently are highly dangerous. But winds of storm strength generated in this manner occur more often in Keszthely and probably, the mentioned more

Table 20. Characteristic values of various storm types in Sicfok

	F	FBo	FBe	Bp	Bs	Be	Bg	N	S
1 Proz. Verteilung der starken Winde	12	21	1	8	17	7	4	24	6
2 Proz. Verteilung der Stürme	12	31	3	3	18	9	3	9	12
3 Mittlere Dauer der starken Winde	1,5 St.	6,8 St.	3,1 St.	1,2 St.	5,5 St.	11,6 St.	5,7 St.	13 Min.	24 Min.
4 Mittlere Dauer der Stürme	1,3 St.	4,6 St.	4 Min.	8 Min.	2,8 St.	4,2 St.	3,0 St.	2 Min.	7 Min.
5 Windrichtungsschwankungen im Laufe des Sturmes	26°	27°	10°	20°	20°	20°	20°	12°	50°
6 Schwankungen der Windstösse im Laufe des Sturmes in m/s	10	7	5	6	5	6	12	2	7
7 Abflauen des Windes von 15 m/s auf 10 m/s	1,1 St.	2,3 St.	1,9 St.	48 Min.	2,5 St.	2,1 St.	2,5 St.	14 Min.	19 Min

Key: 1 percentage distribution of strong winds; 2 - percentabe distribution of storms; 3 - average duration of strong winds; 4 - average duration of storms; 5 - fluctuations in wind direction in the course of the storm; 6 - fluctuations in wind gusts in the cour course of the storm, in m/s; 7 - decline of the wind from 15 m/s to 10 m/s

Table 21. Characteristic values of various storm types in Keszthely

	F	*FBa*	*FBo*	*Bp*	*Ba*	*Bc*	*Bg*	*K*	*S*
1 Proz. Verteilung der starken Winde	20	15	1	8	18	5	11	20	2
2 Proz. Verteilung der Stürme	15	20	2	2	22	4	9	22	4
3 Mittlere Dauer der starken Winde	1,3 St.	8,1 St.	22,0 St.	1,1 St.	7,4 St.	8,8 St.	3,0 St.	7 Min.	21 Min.
4 Mittlere Dauer der Stürme	51 Min.	7,3 St.	20,8 St.	7 Min.	6,4 St.	7,8 St.	1,3 St.	4 Min.	5 Min.
5 Windrichtungsschwankungen im Laufe des Sturmes	24°	28°	45°	20°	21°	45°	52°	11°	23°
6 Schwankungen der Windstösse im Laufe des Sturmes in m/s	13	14	18	12	14	15	3	11	15
7 Abflauen des Windes von 15 m/s auf 10 m/s	1,8 St.	3,5 St.	4,5 St.	20 Min.	2,6 St.	4,3 St.	1,8 St.	15 Min.	5 Min.

Key: 1 - percentage distribution of strong winds; 2 - percentage distribution of storms; 3 - average duration of strong winds; 4 - average duration of storms; 5 - fluctuations in wind direction in the course of the storm; 6 - fl uations in wind gusts in the course of the storm min m/s; 7 - decline of the wind from 15 to 10 m/s.

turbulent structure of the wind there plays a role. In Siófok, most storms are connected with the passage of a cold front, whereby a wedge with nearly West-East axis is formed from the anticyclone located behind the front (frequently lying over the Azores)(situation FBa). At the edge of the anticyclonal wedge, powerful and often long-lasting North storms occur in the Transdanube basin. If this penetration of the Western anticyclone is not accompanied by a sharply defined front, then we have situation Ba, which is likewise at the peak of storm causes in the Lake Balaton area. Under consideration that the origination of the squall line structure occurs very often in the southwestern part of the Transdanube basin, and that it usually moves Northeast, the situation results that well-defined squall line storms with strong winds are three times more frequent over the northeast basin of Lake Balaton, than over the Keszthely area.

The convective storms are the shortest; for a large number of them, the wind gusts reach the threshold value only once or twice. Usually those storms connected not only with the passage of an atmospheric disturbance, but also with a baric system, are longer-lasting. A strong pressure gradient can exist in a region for a long time, consequently the baric storms sometimes last several days. The same also applies for the decline of wind after the storms: The frontal storms, convective storms and squall-line storms quickly loose their strength after passage of the particular atmospheric structure; but the decline of the baric gradient is often a very slow process and the decline of a baric wind takes several hours on average.

With regard to the duration of storms and to the time needed for a decline in the wind, no characteristic differences could be found between the two basins of the lake for the various storm types. Of course, this does not mean that such differences do not exist, but in order to detect them it would be necessary to analyze either the individual cases, or--in the interest of drawing more generalized conclusions--to select a much longer study period as a basis of the work.

Essentially, the same comments about fluctuations in wind speed and direction, occurring during storms, are valid. To be sure, the data from the study period in 1963 were included in the particular tables, but there is no opportunity to draw more general conclusions than those presented here.

2.5 Frequency Distribution of Storm Wind Gusts as a Function of the Average Wind Velocity (P. Ambrózy)

As was already shown in sections 2.3 and 2.4, in the wind conditions of Siófok and Keszthely there are sometimes considerable differences. In addition, there is a factor which is always present--at least in the case of a Northerly wind--for the more lively winds in Keszthely, and this is the gusty nature of the wind. Since the wind recording instrument in Keszthely is rather far from the water surface, no friction-reducing effect occurs, and even for relatively low wind speeds, the formation of important gusts will occur.

With regard to an increase in accuracy of storm warnings it is of interest to determine the frequency and direction of wind gusts, their dependence on the average wind speed, for the two observation posts, i.e. for Siófok and Keszthely. As a threshold value for wind gusts, a speed of 15 m/s was chosen as the basis for the occurrance of a storm state. Naturally, storms beginning suddenly and lasting a short time were excluded from this survey, i.e. cases connected with thunderstorm activity or with an instability line, since a completely different method is needed to investigate them. Practically, this means that the prefrontal and post-frontal winds, and winds occurring in strong baric gradients, are included in the processing.

As starting material we used the hourly values of wind speed exceeding 4 m/s in Keszthely and Siófok from 1 May-30 Sept. for the three years 1961 to 1963. For Keszthely there are 1959 bits of data, and for Siófok, 3876. Classification by wind direction gives:

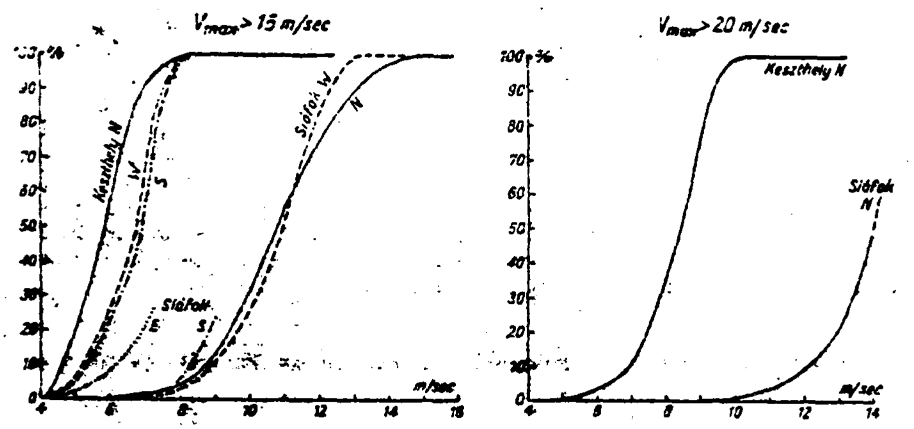
	N	E	S	W
Keszthely	1572	73	220	94
Siófok	2210	93	353	1221

It was assumed--and the validity of this assumption was apparently justified by the processing--that for the introduced limitation (i.e. excluding cases connected with thunderstorms) at an average speed of less than 4 m/s, practically no wind gusts of 15 m/s or more will occur.

In Keszthely, a wind gust exceeding 15 m/s occurred in 44% of all cases, in Siófok, the figure was 11%. A breakdown to wind direction gives:

	N	E	S	W
Keszthely	51	15	14	23 %
Siófok	16	11	3	4 %

This distribution reflects the known fact that the North wind has a greater speed at both places than winds from the other directions.



Figs. 8-9: Frequency Distribution of Wind Gusts Exceeding 15 m/s or 20 m/s whose Hourly Values of Wind Velocity for Different Wind Directions, for instance, for Norverd Winds [Omission in the Text]

The frequency distribution of hourly values connected with wind gusts exceeding 15 m/s, is shown for both places in fig. 8. Due to the small number of cases, a frequency curve for the East wind was omitted, and some other curves are only outlined for the area due to the insufficient number of observations. When viewing these figures, it is immediately striking how far the W and N curves for Siófok are from those for Keszthely, i.e. in Siófok the wind gusts of 15 m/s begin in the case of W and E wind directions, only at much higher values of average wind velocity. In Keszthely, average

velocities greater than 8 m/s are accompanied in all cases by wind gusts of more than 15 m/s, but in Siófok, the same average velocities are connected with wind gusts above 15 m/s in only 4-6% of cases. The cause of this prominent difference can be found in the orographic conditions of the direct vicinity of the observation posts. In Siófok the directions from the semi-circle WSW-ENE indicate a landward wind and consequently, only for high values of average wind-gust velocity will gusts over 15 m/s occur. In the case of a southerly wind direction (SSE to SW), the curve is shifted to the left, and for Easterly winds, the curve is similar to that of Keszthely.

The high frequency values found for Keszthely do not mean that the number of stormy wind gusts is several times greater there than in Siófok (there is only half as much base data for Keszthely as for Siófok). In the course of the three years, in Siófok there were 418, in Keszthely 859 cases of an hourly average accompanied by a wind gust of more than 15 m/s; thus the ratio is only 1:2. At any rate, this is a notable amount.

Similar frequency curves are obtained if a speed of 20 m/s is taken as threshold value. Here, the number of cases is much smaller (241 for Keszthely, 56 for Siófok), but from fig. 9 it is visible that the curves constructed for North winds have only a slight shift to the right, compared to curves in fig. 8, and the difference between the two observation posts is unaffected.

From this investigation, the following conclusions are drawn for the Storm Warning Service and for Weather Prediction:

1. If the average wind velocity in Keszthely exceeds a value of 6 m/s, or if a condition is expected in which a value of 6 m/s will be exceeded, then it must be expected (mainly for a North wind direction) that in more than 90% of cases, wind gusts of storm strength (i.e. more than 15 m/s) will occur. If the average velocity in Keszthely reaches a value of 8 m/s, then the probability of wind gusts of 20 m/s is increased (to more than 70%).

2. In Siófok for landward winds at an average speed of 12 m/s,

and for offshore winds at an average speed of 8-10 m/s, there is a great probability (80%) of wind gusts of storm strength.

2.6 The Wind-Shadow Effect of the Balaton Upland (E. Titkos)

The sailboats on Lake Balaton frequently rely on the Storm Warning Service in Siófok for information on present and expected wind conditions on the lake. In the case of a northerly wind direction, under all circumstances, the wind protection effect must be taken into account; this effect is exerted by the orographic structure of the North bank of the Lake. But there has been no information available about the magnitude of this effect. Consequently, the employees of the Storm Warning Service decided in the course of the summer 1962, to set up a cup-type anemometer on a passenger ship regularly passing between Siófok and Balatonfüred, and to take wind measurements along this route across the Lake in a Northwest-Southeast direction (fig. 10) in any weather situation, where the wind direction is Northwesterly and consequently a wind-shadow might occur along this route.

During the measurements, the reading of the anemometer was taken at intervals and noted together with the time of reading. Since the speed of the ship was known, by using the equation G.Z=S, where G is the speed of the ship, Z the time since casting off and S the distance covered, the location of the ship can be computed at the time of the reading. Thus it was possible to enter all measurement points along the route on a map.

Fig. 10. Route of the Ship Equipped with Anemometer

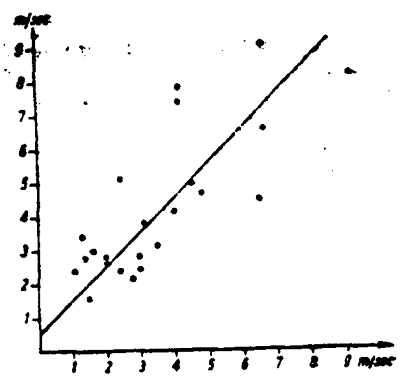


Fig. 11. Distribution of Wind Velocities along the Route Siófok--Balatonfüred

In the course of the data processing, not only the observed wind speeds were entered on the map, but those recorded simultaneously in Siófok were compared and the percentage decrease from the value at Siófok was computed. These percentages were then represented on the map, as seen in fig. 11.

From this figure we see that when moving from Siófok in the direction of Balatonfüred, in the course of the first 3 km, a constant wind speed is observed, but then a slow, and later rapid decrease in wind speed occurs, and in Balatonfüred, the speed is only 60% of that in Siófok.

At the same time, it is supposed that the fluctuations in wind velocity when approaching the North bank of the lake undergo an increase, and from fig. 11 we see that the scattering of points representing the percentage values of wind speed, shows an increase in this direction, which can be interpreted as a consequence of the turbulence generated by the orography.

3. Methods for Predicting Wind on Lake Balaton

3.1 Introduction (G. Tänczer)

This section is concerned with one of the most difficult and important practical questions of the Storm Warning Service at Lake Balaton, namely the prediction of wind. This problem is composed of two tasks whose solutions are based on two completely different methods. First, the wind generated by large-area air-pressure fields must be predicted, with a view toward local conditions. Then, independently of the first task, the temporary wind increases must be predicted, which are caused by synoptic objects of mesosynoptic magnitude (convective thunderstorms, instability lines); these have a relatively short duration and do not extend over large areas.

With regard to the first problem, the suggested method is an investigation of the relation of the wind and the pressure gradient. The prognostic utility of this relation--based on the assumption, reinforced by our investigations, that the present wind is determined by an earlier value of the air pressure gradient--is outlined. This result also supports the Ertel description over that

of Brunt and Douglas. A prediction of the wind originating from the large-scale air-pressure distribution for several hours, and even for half a day, is possible through the method of Litvinova. According to our investigations in this regard, this method can also be used to predict the wind at Lake Balaton.

The second task consisting in the prediction of wind increases connected with thunderstorms, leads to a derivation of the speed of the max. wind gust from the high-altitude air layer. Observations indicate that there is a close relation between the intensity of the wind gust and the near-ground temperature distribution, and that the gust can be computed on the basis of the moisture-potential temperature in the Cumulus condensation layer.

3.2 Attempt to Predict Wind in the Balaton Area on the Basis of the Air Pressure Field (P. Ambrózy, G. Koppány and T. Tänzcer)

Among all meteorologic elements, perhaps the wind speed is that element having the greatest spatial and chronological variability. This is a consequence of the turbulence of air movements. For this reason alone, it can be expected that the prediction of wind parameters (direction, speed, route, gusts) will not be an easy task.

Basically, we are facing different problems according to whether one wants wind predictions for a macrosynoptic area, or for a smaller area, e.g. for the Lake Balaton area. A functional relation is known to exist between the wind speed and the baric gradient, which is specified by the equations of motion. But these relations cannot be used directly in the prognostication. The hydrodynamic methods which serve for predicting the pressure field and wind field, provide only rough wind preditions which are valid only for the macrosynoptic area. They contain no modifying influences of constant character having a mesosynoptic or microsynoptic extent (orography, roughness of the surface), and also contain no wind fluctuations occurring during the prediction period due to mesosynoptic and microsynoptic objects (a thundercloud, an instability line etc.). It should be noted that when predicting wind, the wind strength is more important

than the wind direction. As regards the wind direction, it can be determined with sufficient accuracy by means of the pressure field.

In agreement with the foregoing, we find that in the prediction of wind conditions of Lake Balaton, we can rely only somewhat on the general, large-area predictions, since such predictions must be modified by the finer structure of the pressure pattern and the influences of the environment.

Within the frame of the prediction of wind conditions in the Balaton region, the following problems come up:

1. It is necessary to come up with a general prediction for 12 hours for the lake region. This prediction is produced on the basis of knowledge of the large-area pressure and wind distribution, by extrapolation. When citing wind direction and speed, no other details are presented than are published in the wind predictions for the entire region.

2. It is necessary to obtain a more accurate knowledge of the structure of wind speed for a period of 2 to 3 hours within the period of validity of the prediction, as is the case in the general prediction (with an error limit of $\pm$ 2 m/s). This is an indespensible condition for a timely issuance of storm warnings.

3. The Storm Warning Service should also issue other, short-term predictions for sailing matches, boat tours etc., whereby local circumstances are taken into account.

The problems mentioned in 2 and 3 above, are essentially identical, since both are based on a relatively accurate prediction of wind speed valid for only a few hours.

To approach this question, one may proceed from the geostrophic relation resulting from the hydrodynamic equations. They of course, possess no prognostic value, but they do reflect the relation between the pressure gradient and the geostrophic winds. In

reality, the wind does deviate from the geostrophic wind, and one of the problems is to determine this deviation. Since the wind in every case is braked by friction, in the majority of cases we have a wind speed less than the geostrophic wind. [Misprint in the text...]to determine the relation existing between the baric gradient and the actual wind speed, by statistic means. Ertel [33] has demonstrated that the structure of the wind follows the change in gradient with a time delay, i.e. a so-called adaptation time occurs. This information has a prognostic significance. Thus, the hope exists that the present value of the baric gradient can be used to predict wind conditions for one to two hours in advance.

Lastly, we arrived at the solution that in a statistic manner, a relation between the baric gradient and the actual wind speed can be sought, and on the basis of this relationship, we shall draw the necessary conclusions for establishing a prediction with reference to the Ertel description. Two experiments in this regard are presented; both are concerned with wind predictions for Siófok.

3.2.1 Determination of Wind Velocity Based on Air Pressure Data from Two Observation Points

In Siófok the northern winds have the greatest frequency. This fact is particularly true for the strong and stormy winds. Consequently, our investigation was limited exclusively to winds having a Northerly component, and we searched for the relations of these winds with the pressure gradient. With respect to the gustiness of the wind, we considered it expedient to use the hourly wind travel instead of the instantaneous values of wind speed, in this experiment. Since the Northern winds in Siófok are landward winds, the fractional influence on the wind is less than for seaward winds. Thus, we can postulate that the northern winds are generated by a nearly West-East directed air-pressure gradient. Accordingly, the wind travel in Siófok can be related to the pressure data of the synoptic observation posts of Szombathely and Kecskemét, e.g. the wind travel occurring between 12 and 13 hours can be compared to the pressure difference from the synoptic observations at 1240 hours. The connecting line between the two observation posts forms

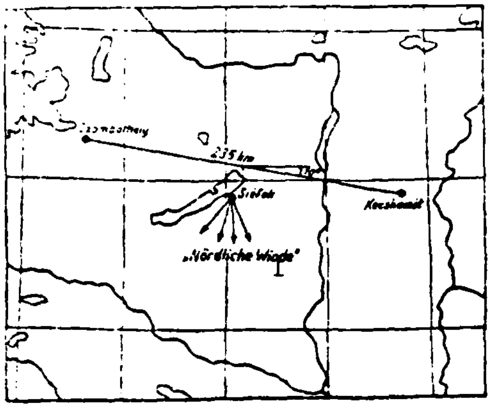


Fig. 12. Location of Stations Used to Compute the Northern Winds of Siófok.

Key: 1-Northern winds

an angle of 10^{o} with the circle of latitude. For this reason, winds from the directions NNW, N, NNE and NE were counted among the Northern winds (fig. 12).

As base material for this investigation, we used data from the period of 1 May-30 Sept. from the years 1961-1963. In order to avoid disturbances due to local circulation of Lake Balaton and other microsynoptic and mesosynoptic effects, the winds weaker than 15 km/hour were eliminated from the study. To investigate the isollobar influence, those cases in which an isallobar difference of more than 1.0 mb/3 hours occurred at the observation posts, were sorted out. The investigation of stability conditions was performed by means of the stability coefficients obtained by Szepesi [34] for the ground air-layer; these were computed from the Budapest Radioprobe Ascent. The cases sorted out were entered into a coordinate system whose abscissa denotes the Szombathely--Kecskemét pressure difference and whose ordinate represents the one-hour wind travel in Siófok. The relation between these two quantities was evaluated in such a manner that we split the deviations from the line--represented by the geostrophic wind--into intervals of 6 km/hour each, and the number of cases for each such interval was determined below and above the mentioned line. The series of curves used as a basis for this evaluation is presented in fig. 13. Table 22 contains the percentage

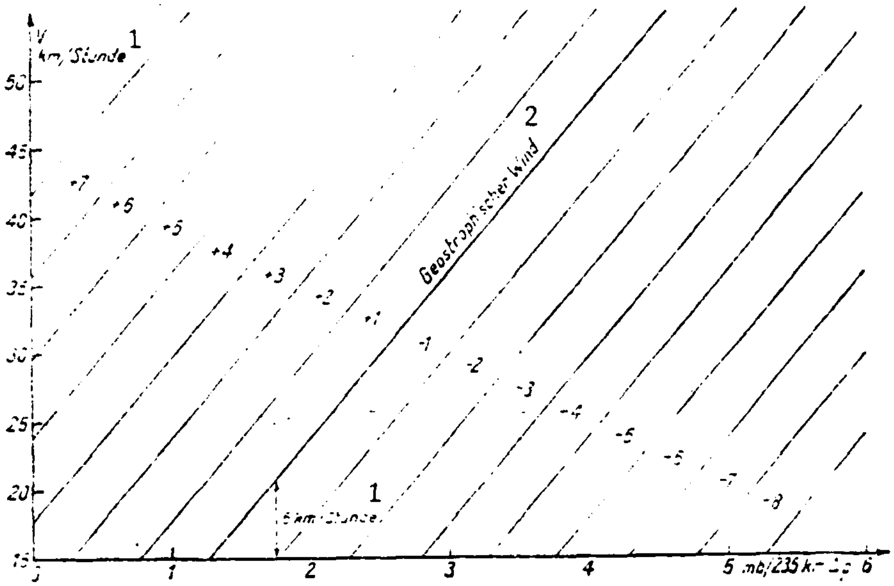


Fig. 13. Family of Curves to Evaluate the Relationship Existing Between the Pressure Gradient and the Wind

Key: 1-hour 2-geostrophic wind

Table 22. Deviations of the Wind from Geostrophic Winds for Northerly Wind Directions (Z=Number of cases)

Year	Z	−6	+5	−4	+3	+2	−1	−1	−2	−3	−4	−5	−6	−7	−8	C
1961	582	0	0	0	0	2	4	7	18	28	21	11	5	[illegible]	1	67
1962	651	1	1	3	6	10	17	21	16	11	7	4	2	[illegible]	0	54
1963	556	1	3	2	9	17	24	18	15	6	3	1	1	[illegible]	0	59

distribution of cases with respect to the geostrophic wind.

The numerical value in the third column of the table (C) gives the following information: The number of cases belonging to the three intervals having the greatest frequencies, was found and expressed in percent of all cases n. The total width of the three intervals apparently makes up 18 km/hour (i.e. 5 m/s), and thus represents a tolerance which seems acceptable in this case. It is striking that in 1961, the three intervals of max. frequency were about 12-18 km/hour higher than in 1962 and 1963. The reason for this lies in the fact that in Kecskemét on 1 April 1962, the barometer was replaced which resulted in an increase in pressure values of more than 1 mb, and thus a corresponding reduction in the

Szombathely--Kecskemét pressure difference. To eliminate this system error, the same investigation was also performed for South winds (SSE, S, SSW, SW). The distribution of cases is shown in table 23.

According to this data, the South ("S") winds should be about 5 m/s stronger for the same value of the pressure gradient, than the northern ("N") winds, which is naturally an impossible result. The frequency distribution of the "N" and "S" winds in 1961 compared to the geostrophic winds is illustrated in fig. 14. By adding the two curves, we get the result that the actual wind in Siófok amounts to 70-75% of the geostrophic wind.

Table 23. Deviations of the Wind from Geostrophic Wind for Southerly Wind Directions (Z=Number of cases)

Year	Z	+6	+5	−4	+3	+2	+1	−1	−2	−3	−4	−5	−6	−7	−8	C
1961	121	1	2	9	12	19	21	21	12	3	1	0	0	0	0	61

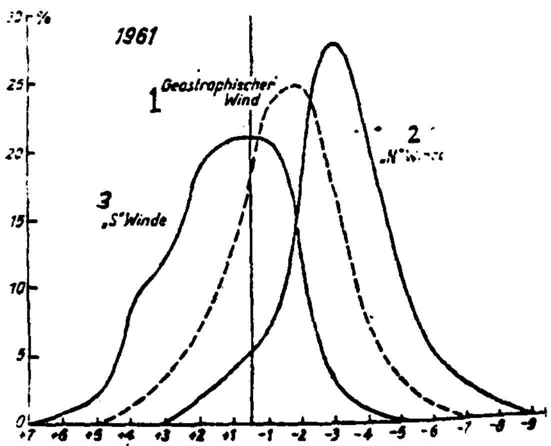


Fig. 14. Frequency Distribution of the Northerly Winds in 1961. The Dashed Curve Represents the Averages Formed from the Values "N" and "S".

Key: 1-geostrophic wind 2-N-winds 3-S-winds

In the distribution for the years 1962 and 1963, we find an opposite situation; here we encounter gradients smaller than the actual gradient value. Except for these, the data in all three years exhibit a rather large scattering. It turns out that winds occur which are much stronger or weaker than the geostrophic wind; thus

the frequency of the latter is greater. Whereas the excessively strong winds are explained mainly by thunderstorm activity and frontal passages, the ground inversions must be responsible for the weak winds.

In the introduction it was mentioned that Ertel says the development of the pressure field preceeds the structure of the wind, i.e. the instantaneous values of the wind are defined by a state of the pressure field occurring earlier. For this reason, in 1961 we examined the gradient values occurring 1, 2 and 3 hours earlier, and one hour later--in addition to the simultaneous figures. The results of this investigation are presented in the same manner as in table 23 (table 24).

To be sure, in the 'C'-values--which describe the closeness of the existing relations--no significant differences result; but the best relationship occurred in those cases where the one-hour wind travel was linked with the gradient observed an hour earlier. This result seems to support the Ertel conception, where the adaptation time should be about one hour. The fact that the relation between the instantaneous pressure gradient and the wind occurring 3 hours later is only 6% worse, justifies us in making a 3-hour wind prediction on the basis of the present pressure field. Naturally, in such cases, one must proceed with all due caution.

Table 24. Deviations of the Wind from Geostrophic Wind by Using Time-Offset Values of the Pressure Gradient (Z=Number of cases)

Year	Δt	Z	+4	+3	+2	+1	−1	−2	−3	−4	−5	−6	−7	−8	C
1961	−3	527	0	1	3	3	8	16	28	19	13	6	2	1	63
1961	−2	588	0	1	2	4	6	21	23	21	12	6	2	1	65
1961	−1	571	0	0	1	3	7	21	26	22	12	6	2	1	69
1961	~0	582	0	0	2	4	7	18	29	21	11	5	3	1	67
1961	+1	521	0	1	3	6	12	20	26	17	8	4	2	1	63

3.2.2 Investigation of the Isallobar Influence

As already mentioned, those cases where an isallobar difference of more than 1.0 mb existed in the 3-hour trend values between Szombathely and Kecskemét, were sorted out. The extent to which this

sorting affected the distribution, was examined. The result is presented in table 25.

Table 25. Deviation of the Wind from the Geostrophic Wind, Excluding Isallobar Cases [illegible] hours]. (Z=Number of cases)

Year	Z	[illegible]	-5	-4	3	2	1	-1	-2	[illegible]	-4	[illegible]	[illegible]	-7	-8	[illegible]
1961	[illegible]	0	0	0	0	2	2	7	[illegible]	[illegible]	22	11	[illegible]	3	0	71
1962	[illegible]	1	1	3	6	9	17	22	[illegible]	12	[illegible]	2	2	1	0	57
1963	428	0	3	1	10	18	24	18	[illegible]		3	1	1	0	0	60

From the table we see that through the exclusion of those particular cases in all three years, an improvement in the relationship can be achieved, even though this improvement is not a significant one. Since the isallobar influence exerted an unfavorable effect on the relation between air-pressure gradient and wind travel, it seems useful to examine cases which deviate from the average distribution. It must be determined whether any regularities are present in the distribution, either by the size or by the sign of the isallobar gradient. The number of points was determined which are located above and below the three intervals of max. frequency, i.e. a determination of the number of cases in which the wind strength was greater or smaller than the average speed, took place. The results are reported in table 26 with a differentiation of cases with positive and negative isallobar difference. An isallobar difference was considered positive, even when the difference of trend values from Szombathely and Kecskemét has a positive sign.

Table 26. Deviations of the Wind from the Geostrophic Wind due to the Isallobar Effect (Z=Number of cases)

Year	Z	$\Delta\,\partial p/\partial t < 0$			Z	$\Delta\,\partial p/\partial t > 0$		
		$v > \bar{v}$	$v \sim \bar{v}$	$v < \bar{v}$		$v > \bar{v}$	$v \sim \bar{v}$	$v < \bar{v}$
1961	39	17	19	3	46	5	18	23
1962	41	18	19	4	73	13	29	31
1963	62	15	34	13	69	6	35	28
Σ	142	50	72	20	188	24	82	82
%	100	35	51	14	100	13	44	43

From the table we see the surprising result than in those cases of a negative isallobar gradient, an above-average wind speed occurs, but in cases with positive isallobar gradient, below-average

winds occur. It should be noted that the examined isallobar differences do not necessarily coincide with the isallobar gradients, since they are actually only the West-East component of this gradient. At any rate, these quantities are sufficient for demonstrating the isallobar influence. The behavior of cases in which an isallobar influence is present, is quite prominent if we note that it applies to only about 13-14% of cases. From this fact there can be no doubt that a regular, recurring event is being made known. In order to discover this regularity, let us turn to an analysis of the concept of the isallobar wind.

As is known from the Brunt-Douglas interpretation of the isallobar wind, the following relation exists:

$$\mathbf{v}_{is} = -\frac{1}{\varrho f} \nabla_H \frac{\partial p}{\partial t}, \tag{3.2.1}$$

from which follows that the direction of this quantity must coincide with the direction of the isallobar gradient. In fig. 15, the influence of the West-East isallobar gradient on the wind in Siófok is shown. We see that this influence consists mainly in a change in wind direction and an increase in its strength. A decrease in wind strength could only be caused by a South-to-North running isallobar gradient; in such a case, the isallobar difference of Szombathely--Kecskemét would nearly disappear. Our studies are thus not in harmony with these conclusions, since we had to admit that precisely an opposite effect is present. Thus arose the need to find another theoretical explanation for the obtained distribution.

Now in the technical literature, another interpretation of the isallobar wind is known, namely that of Ertel, which is based on the fact that the wind is determined by the geostrophic wind belonging to an earlier time, i.e.

$$\mathbf{v} = \mathbf{v}_{g_0}. \tag{3.2.2}$$

However, the instantaneous value of the geostrophic wind can be written as follows:

$$\mathbf{v}_g = \mathbf{v}_{g_0} + \frac{\partial \mathbf{v}_g}{\partial t}(t - t_0), \tag{3.2.3}$$

from which follows:

$$\mathbf{V}_{g_0} = \mathbf{V}_g - \frac{\partial \mathbf{V}_g}{\partial t}(t-t_0). \tag{3.2.4}$$

If this expression is substituted into (3.2.2), we have:

$$\mathbf{V} = \mathbf{V}_g - \frac{\partial \mathbf{V}_g}{\partial t}(t-t_0) \tag{3.2.5}$$

and since:

$$\frac{\partial \mathbf{V}_g}{\partial t} = \frac{1}{\varrho f}\nabla_H \frac{\partial p}{\partial t} \times \mathbf{k} \tag{3.2.6}$$

we have the final result:

$$\mathbf{V} = \mathbf{V}_g - \frac{1}{\varrho f}\nabla_H \frac{\partial p}{\partial t} \times \mathbf{k}\,(t-t_0). \tag{3.2.7}$$

Now it is immediately evident that the Ertel value of the isallobar wind is given by the second term on the right side of equation (3.2.7). The expression $t-t_0$ corresponds to the adaptation time at which the wind follows the changes in the pressure field. The isallobar wind interpreted in this manner is directed parallel to the isobars and has a direction lying to the left of the isallobar gradient. This effect is illustrated in fig. 16 for the case of a West-East directed isallobar gradient for a North wind in Siófok.

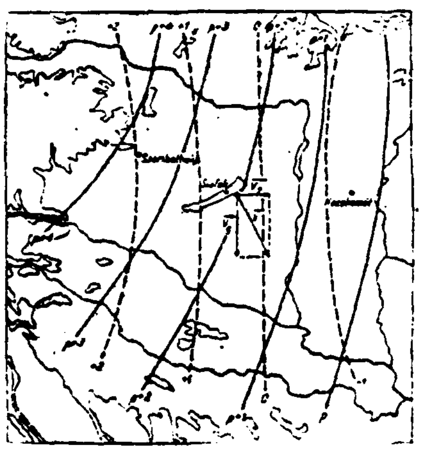


Fig. 15. Interpretation of the Isallobar Wind by Brunt and Douglas

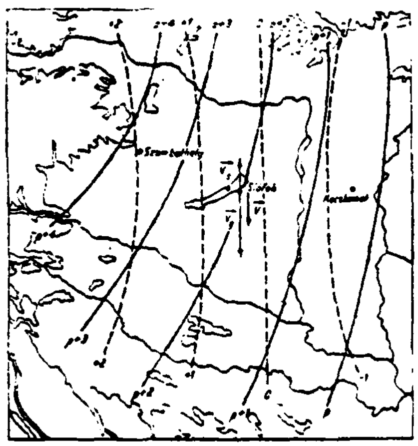


Fig. 16. Interpretation of the Isallobar Wind according to Ertel

Thus a West-East directed isallobar gradient has an effect

exactly opposite to the North wind, i.e. the effect consists in a weakening of the wind and conversely, an East-West directed isallobar gradient leads to an intensification of the North wind. If this conclusion is compared with the results of observations in Siófok, then we find complete agreement. For this reason, it can be postulated with great probability, that the "regularity" found for the isallobar wind might be a result of the effect of the Ertel isallobar wind. This conception is also supported by the fact that the best relationship between wind and pressure gradient can be obtained in the case where the wind conditions were compared with a pressure field which existed one hour earlier.

We have computed the wind velocity which corresponds to an isallobar gradient of 1 mb/3 hours between Szombathely and Kecskemét. We obtained a result of 0.9 m/s. But the computed value is much too small, since in reality even ageostrophic deviations of 5 m/s have been observed. This deviation can have several causes: 1. Incorrect pressure and trend data which may have occurred due to micro- or mesosynoptic processes, or due to incorrect observations; 2. The pressure and isallobar differences between Szombathely and Kecsekmét do not determine the gradient value for Siófok, even though a North wind is doubtless under the influence of the West-East component of this gradient; 3. The adaptation time can be more than one hour, and for instance, for an adaptation time of 1½ hours, one has an isallobar wind which is one and one-half times greater than before; 4. It would be possible that it is more correct to work with a pressure change related to a shorter time than the 3-hour air-pressure trend, since we only used the 3-hour trend for our own convenience. These possibilities may be why it was possible to find a qualitative, but not a quantitative, relation between the isallobar gradient and the isallobar wind.

In order to describe the isallobar influence more accurately, the following example will help. Let us take a wind which comes from the North the entire time it is blowing. Insert a pressure increase from the West; then an increase in the baric gradient and an intensification of the wind must be expected. The isallobar wind component acts in the opposite direction and leads to a reduction

of the wind increase. Consequently, as long as the West-East directed isallobar gradient continues, the wind will be weaker than expected. If the isallobar gradient assumes an opposite direction, i.e. if there is a pressure drop in the West and an increase in the East, then we obtain an opposite isallobar action; the decline in the wind is delayed by this effect and a wind can be generated which is stronger than the geostrophic wind. To illustrate this process, we introduce fig. 17. In the upper part of the figure, the change in pressure gradient is shown. The middle curve shows the change in isallobar gradient. Below this is the change in the wind which reflects the configuration of the pressure field with a 1-hour delay. The dashed curve would apply in cases where the wind is adapted to the instantaneous pressure gradient.

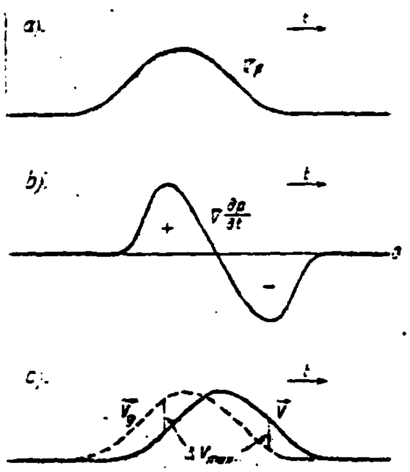


Fig. 17. The Isallobar Effect in a North Wind after Passage of a Pressure Wave

a) Change in the pressure gradient

b) Change in the isallobar gradient

c) Change in the wind

3.2.3 Determination of Wind Velocity based on Calculations of the Gradient at a Circle-Periphery

As we discovered in sections 3.2.1 and 3.2.2, the attempt to find a relation between the pressure difference existing between two fixed points (station pair) and the average speed of the wind, did not lead to the expected result due to the relatively severe scattering of data. The useable conclusions are mainly of a qualitative, and not quantitative, nature. The reasons for this failure were as follows:

1. The pressure difference of two observation posts (Szombathely, Kecskemét) was used not only for wind directions perpendicular to the connecting line between the two posts, but deviations up to 35° from this line were permitted.

2. It is by no means certain that the Szombathely--Kecskemét pressure difference corresponded to the max. difference for each wind direction. In fact, only the gradients with directions between SE and E are considered correct (which correspond nearly to the direction of Szombathely--Kecskemét); thus in 18 percent of all cases, this condition is not met.

3. By using a single pair of stations, only for some directions of the wind rose can relatively reliable pressure differences be computed, and for all possible wind directions it would be necessary to use at least 4 pairs of stations. But within the existing station network, no four pairs of stations can be found which are at the same distance from Lake Balaton, along the four primary and four secondary sky-directions and which have hourly observations.

4. If a given station pair or if given station pairs are used, while ignoring the data from surrounding stations, then there is no possibility of correcting errors occurring in the course of observations, air-pressure reduction and transmittal of messages. In addition, as discussed in section 3.2.1, a replacement of instruments can cause former relationships to be changed entirely.

The above incongruities can be eliminated if, instead of station pairs, the following method is used: Around Lake Balaton, or more accurately, around Siófok, draw a circle of appropriate size, at whose periphery the greatest and smallest pressure value is drawn in by interpolation; thus also the direction and size of the air pressure gradient is included (fig. 18). The radius of the circle is selected as 150 km. This value is large enough for a sufficiently accurate determination of the direction and size of the gradient, and at the same time, has the added advantage that the circle hardly goes beyond our national borders and if necessary, one can make do with domestic pressure observations. The determinaiton of the gradient was performed exclusively on the basis of pressure data; no wind parameters were taken into account.

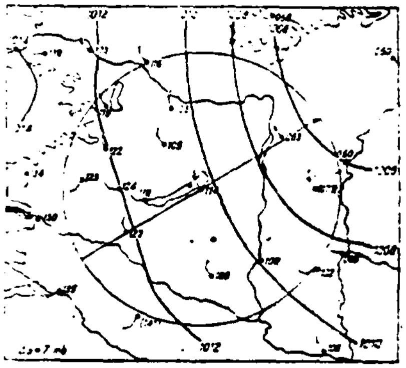


Fig. 18. Determination of the Pressure Gradient (22 August 1963, 06 hours, Greenwich Mean Time)

The computed values of the gradient were related to the wind velocities in the following manner. The gradient value related to a specific time, was brought into harmony with the max. hourly average of wind speed from the next three hours as per the above conception of the "adaptation" of the wind (i.e. for the three hourly averages, the greatest value was selected). As will be shown in a later summary, this method leads to the best results.

It can be presumed that for different wind directions, different relations result between gradient and wind, but the deviations will not be significant--if there are no differences in the orographic conditions within a circle of 5-10 km radius about the observation post. But the situation at Lake Balaton is quite different; here, two basically different types of earth's surface come together. When using wind data from the Siófok observatory, the large differences between landward and seaward winds are immediately obvious. This effect is more pronounced than the differences resulting from different wind directions under identical orographic circumstances.

The relation existing between the air-pressure differences on the one hand (determined by means of the above method of gradient determination based on pressure values on a circle) and the wind speeds on the other, was presented in fig. 19 for the summer of 1964 in Siófok. As we see, the values denoted by a large dot (relating to offshore winds) are located in the lower part of the pile of dots. This means that for the same values of the air pressure gradient, the offshore winds are usually weaker than the landward ones. In the figure, the geostrophic wind speed is shown as a

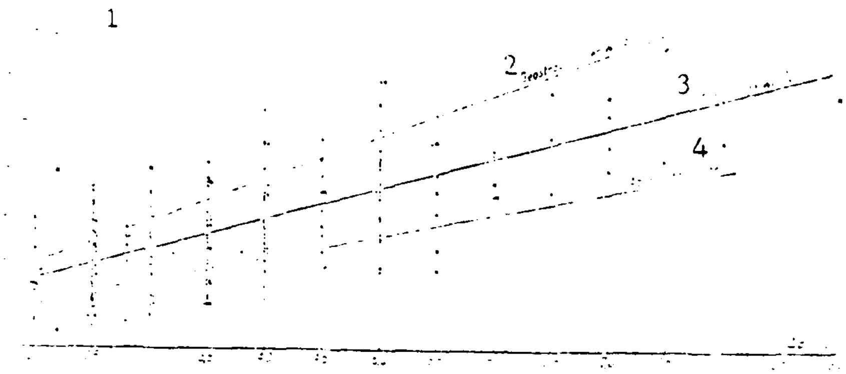


Fig. 19. Relation of Wind Velocity with the Pressure Difference, Siófok, May-September 1964.

Key: 1-km/hour 2-geostrophic wind 3-landward wind 4-offshore winds

broken line, the bold line represents the regression line for landward winds and the thin line is the regression line for offshore winds. The differences in surface are expressed rather well by the distance of these two lines from each other, and by the slow divergence of their values.

In selecting the cases, only three restrictions were imposed: a) Thunderstorm situations were excluded since the intensity of the wind caused by a thunderstorm, is almost independent of the pressure gradient. Section 3.3 is concerned with the prediction of thunderstorm wind gusts.

b) Cases in which the pressure differences along the circle were less than 2 mb were excluded. In situations of weak gradient, the wind speeds are naturally also low. The numeric ratio of cases excluded in this manner, was determined in the following manner. The cases were counted up in which the gradient was less than 2 mb/300 km, but where the wind speed had a value of at least 4 m/s. This amounted to 15-18% of all cases. If we take into account that the majority of these cases consists of winds of 4-5 m/s, then we can assume that only a few percent are due to the higher speeds which are important for wind prediction. From this it follows that this omission contains no danger.

c) The night cases were excluded from the investigation. This means that the decrease in turbulence and wind decrease caused by inversions did not have to be taken into account any further. The simple

explanation for this restriction is due to the fact that when processing, the weather charts were used which were prepared by the Storm Warning Service during the day in 3-hour intervals; but none were available for the night. In addition, note that the processing extends only to the period of activity of the Storm Warning Service (15 May to 30 September); thus seasonal differences could not be taken into account.

As mentioned above, by suppressing the development of the wind field compared to the development of the pressure field, the potential is created for drawing conclusions from the baric pattern existing at a given time, about wind conditions in a later time. As we see in table 27, we obtain no significant deviations between the following comparisons: Premature comparisons (in which the speed data come from the hour before the time of determining the gradient), simultaneous comparisons and "subsequent" comparisons; but the best results are obtained for the maximum, subsequent comparisons in each year. The correlation coefficients included in the table relate to landward winds. It is striking that in 1962, the relation between baric fields and wind is weaker.

Table 27: Relationship Between Pressure Gradient and Wind Speed (Correlation Coefficient with error in parentheses)

	max. subsequent	simultaneous	premature
1962	0,61 (0,03)	0,60 (0,03)	0,59 (0,03)
1963	0,72 (0,02)	0,64 (0,03)	— —
1964	0,72 (0,02)	0,66 (0,02)	0,66 (0,02)

For offshore winds, the situation is much more confused. Since the offshore winds make up only 24% of all cases (all winds from a direction of NE to SW are offshore winds, i.e. they account for more than 180° of the horizon) we obtained very low correlation coefficients (0.1 - 0.2) due to the small number of cases and the excessive scattering. The year 1964 was an exception and a value of 0.59 resulted (with an error of 0.05). For the latter case, a regression line in fig. 19 was suggested. Since the regression lines of the three studied years differed only a little, the lines in fig. 19 can be considered to be generally valid for the area of Siófok.

The equations of these lines are:

$$y=2.0+6.7x \quad \text{(landward winds)}$$
$$y=0.3+4.9x \quad \text{(offshore winds)}$$

The loose relation existing between the offshore winds and the baric gradient can be illustrated by specifying the direction deviations between the actual winds and between the geostrophic winds, which is defined by the pressure gradient. From fig. 20 it is visible that the landward wind direction deviates less from the theoretical values (deviations of 0 to 45° have a high frequency), and for offshore winds there is a significant scattering, especially in a positive direction. This apparently acts on the values of wind speed. In both distributions, we see the onesided deviation caused by ground friction (the maximum is located not at 0° but between 0 and 45°). On the basis of an approximation calculation, for the average deviation from the geostrophic wind of landward wind, there results a value of 26° and for offshore winds, a value of 33°.

Direction deviations of more than 90° usually occurred for weak gradients and small wind speeds. In such cases, the land and lake wind generated by temperature differences between water and lake, might exert a much stronger action on the structure of wind direction than does the pressure gradient.

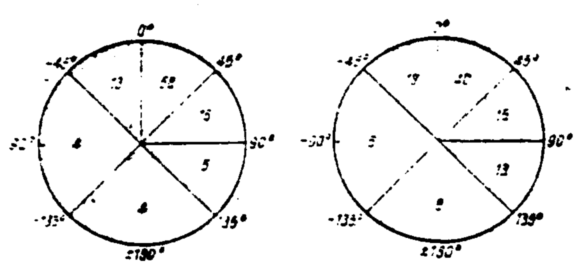


Fig. 20. Deviations between the Directions of the Geostrophic and the Actual Wind (percentage distribution)

3.2.4 Relation to the Weather Situation

By theoretical means it is easy to prove that for an equal value of the air pressure gradient, there are greater wind speeds in anticyclones than in cyclones. Thus the question arises of whether it might be appropriate to perform the above wind prediction in a manner such that takes into account the type of synoptic structures generating the particular gradient. The problem could be solved such that

the relation of the baric gradient with the wind speed is solved separately for each type of weather situation. Another solution consists in determining these relations for the various sectors of cyclones and anticyclones. Within these sectors, distinctions can be made between stable and instable air states. But if all these circumstances are to be specified, then one needs an exceptionally large number of data in order to have a sufficient number of cases for each class. Consequently, we split the cases only so much that the isobar structures were ordered into three types (cyclonal curvature, anticyclonal curvature and indefinite curvature). Since in the determination of curvature, only isobar shapes of pronounced curvature were permitted, 67% of all cases fell into the class of indefinite curvature, and the number of anticyclonal cases was only 19%; the cyclonal cases were only 14%. By excluding cases of indefinite curvature, by comparison of anticyclonal and cyclonal cases (table 28) one fact was certainly proven: That for an anticyclonal curvature, a higher wind speed occurs for equal values of the pressure gradient. The differences resulting between the two curvature types, are reduced slightly since for cyclonal curvature, the fraction of offshore winds (and thus weaker winds) was somewhat greater.

Table 28. Average Wind Speeds ($\bar{V}$) for Cyclonal (Z) and for Anticyclonal (A) isobar Curvature as a Function of the Air Pressure, Siófok, 1964

	Δp	20	25	30	35	40	45	50	55	60	65	70	75	80
A	$\bar{V}$	12	21	21	20	32	35	34	37	32	—	56	67	53
	a Zahl der Fälle	12	23	31	17	12	9	10	3	3	—	1	1	2
Z	$\bar{V}$	13	19	18	18	26	17	31	28	36	35	—	—	—
	a Zahl der Fälle	18	12	18	14	12	4	8	1	2	1	—	—	—

Key: a-number of cases

The splitting into stable and labile air states was performed in the course of processing mentioned in section 3.2.1, but since there is no observation post near Lake Balaton where radioprobe ascents can be carried out, the results are loaded with uncertainly and will not be repeated here.

Due to the tests performed here, the potential was created for working out a simple method for predicting wind which relates to the South bank of Lake Balaton, primarily the environ of Siófok, having a validity of several hours.

On the basis of the three-hour synoptic ground maps, the maximum value of the pressure gradient can be determined (fig. 18), then the attendant value of wind speed expected in the next three hours and linked to the greatest hourly average, can be read off fig. 19. Here we use either the "landward" or "offshore" regression line, depending on the wind direction expected due to the baric situation. If we know the average wind speed, then by means of fig. 8 in sec. 2.5 we can determine the probability of a wind gust exceeding a speed of 15 m/s.

3.3 Relation of Stormy Wind Gusts with Stormy Temperature Jumps (E. Bodolai and B. Bójti)

One of the most important tasks of the Storm Warning Service at Lake Balaton consists in predicting wind strength occurring in thunderstorms. In the course of the last few years, several tests were conducted in the area of objective prediction of winds occurring in connection with thunderstorm activity. The present work can be viewed as a continuation of those efforts.

In the synoptic literature it has been known for a rather long time that between the speed of the wind emitted by a thunderstorm and the temperature decrease accompanied by that storm, there is a specific relation [3,4,5]. Faust [4] derived the following empirical formula for linking the temperature jump and maximum wind gust:

$$V_b - \frac{V_v - V_n}{2} \, [\mathrm{m/sec}] = 2\Delta T \, [^\circ].$$

where the symbols have the following meaning:

V_b Vector of max. wind gust on the ground

V_v Vector of average ground wind before the wind gust

V_n Vector of average ground wind after the wind gust

ΔT Temperature drop on passage of the wind gust.

This equation is said by Faust to be a very rigid relation, which only exists if at the backside of the disturbance connected with the wind gust, a moisture-adiabatic value of the vertical temperature gradient is present. In the work by Faust, a theoretical derivant of the above relation is given, and we will return later to some findings of this derivant.

In the interest of a practical valuation of the relation between the thunderstorm temperature jump and the maximum wind gust, Fawbush and Miller [5] derived a regression equation for both parameters based on a series of observations of 60 terms. The correlation coefficient of this series was 0.86. The temperature jump was determined by means of a method of Brancato, from the vertical distribution of the temperature of a moist thermometer, rather, this quantity was identified with the moist-potential temperature of that point obtained by forming the intersection of the curve of state of the moist temperature with the 0 degree isotherm. According to Petterssen [3], this relation proves useful for predicting thunderstorm wind gusts near the ground, mainly in storms having a speed of more than about 20 m/s.

Due to this work, Ambrózy and Tänczer [6] performed investigations on the utility of the Fawbush-Miller method in Hungary. From tests performed on Hungarian data, it was found that due to the regression curve set up by Fawbush and Miller, much too high values are obtained for maximum wind gusts in Hungary. For this reason, the authors constructed a regression curve between wind maximum and temperature jump on the basis of 20 non-frontal thunderstorms at Lake Balaton, where the wind speed exceeded 15 m/s. Efforts were also made to determine how best to approximate the temperature decrease accompanying the thunderstorm. From these studies came the result that the point whose moist-potential temperature is used, should not be taken from the moist, but from the actual curve of state; and to be sure, this is the point of the curve of state having the temperature 0^{o}.

In 1963 and 1964 the method suggested by Ambrózy and Tänczer was used by the Storm Warning Service at Lake Balaton. The results

were not satisfactory, and at the end of the season in 1964, the director of the Storm Warning Service became convinced that this method would have to be abandoned because the synopticians were still being led off by the excessive values of obtained wind speeds.

In the course of the mesosynoptic investigations of instability lines it turned out that the existance of the questionable relation always came into effect in some form and we believed that we could not eliminate the use of such a close relation in the synoptic work. For this reason we shall bring up this problem in a somewhat different light and attempt to solve it.

The existance of a relatively close relation between the thunderstorm temperature drop and the maximum wind gust is elevated beyond all doubt. So why are the synoptic results still unsatisfactory? The following circumstances can be called into question as sources of error:

1. In the course of previous investigations no distinction was made between convective thunderstorms and thunderstorms generated from an instability line in the derivation of regression equations and in working out the thermodynamic method. (Under the designation "Convective thunderstorms", we mean thunderstorms occurring independently of an instability line or cold front, in a local manner, singly or in groups, without regard to the circumstances of their origin, i.e. whether they be thermal thunderstorms or those generated by a forced convergence or as a result of the structure of the near-ground convergence field). In practice too, the Ambrózy and Tänczer regression curve was used in the same manner for convective thunderstorms and for instability lines. The idea for a distinction of these two structures did not come up, since both phenomena occurred within one and the same air mass. In fact, this coincidence does exist, but the two phenomena differ in their organization into the system of large synoptic processes, due to their spatial and chronological extent. An instability line is a mesosynoptic phenomenon, whereas the convective thunderstorms belong to a group of local phenomena. This distinction exists for these two types of thunderstorm and is reflected by intensity

differences in the storms; and this empirical fact requires no special verification in the view of the synopticians.

2. The regression curve of Ambrózy and Tänczer which was also used by the Storm Warning Service, originated from data coming from observation posts in the Balaton region (from Keszthely and Siófok) and so it turned out that the regression curve only reflected conditions occurring in the two windiest regions of the country. It was demonstrated by preceeding investigations [7,8] that the spatial distribution of max. wind-speeds at an instability line have a great variability within the country. It must be noted that the instability lines have a certain life history consisting of a development stage, a period of greatest development, and a dissipation stage, and even within one and the same stage, there are intensity differences occurring in the course of movement of the line. For this reason, it may not be expected that a prediction built on a single ascent, will give temperature jumps valid for all locations of the entire prediction area with sufficient accuracy. The values delivered by a prediction can only be viewed as averages, whereby under certain rational viewpoints, the quadratic scattering can be taken into account.

3. As a third source of error, we have the prerequisites and methods relating to the prognosticated temperature jump. We will return to this problem later in this paper.

The above sources of error also contain new viewpoints for further investigation. These considerations gave us occassion to examine the relation between thunderstorm temperature drop and maximum wind gusts in the case of instability lines, convective thunderstorms and thunderstorms at cold fronts on the basis of a greater volume of data. The processing of the material was simplified since data was available from a mesosynoptic investigation of 17 instability lines and 37 cold fronts with thunderstorms, from the last 10 years on the particular temperature drop and on the particular value of the max. wind gust for the entire land area. In this manner, in the derivation of the regression equation we eliminated

the error consisting of our use of data coming from observation posts on the Balaton region where the max. winds are usually stronger than the maxima recorded at many other observation posts of the country.

The correlation coefficient and the regression equation for the relation between thunderstorm temperature jump and maximum wind gust was derived from 95 bits of data. Among this data are no bits of information relating to the development stage or dissipation stage of the instability line. For the correlation coefficients of the two parameters, we obtained the value 0.68; the value of the quadratic scattering is 3.3 m/s. In fig. 21, the scattering diagram of the two variables is shown; for the following regression equation, we have:

$$V_{max(ie)} = 13{,}2 + 1{,}1\Delta t. \tag{3.3.1}$$

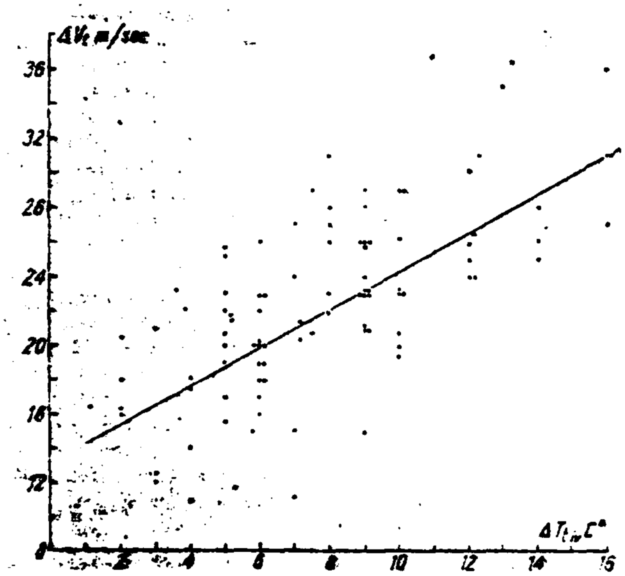


Fig. 21. Relation of Actual Temperature Jumps to the Actual Wind Maxima in the Case of Instability Lines

If this result is compared with that of [6], then it is visible that one will obtain a smaller value for the max. wind gust by using our regression equation, which can be explained by the fact that our data relate to the whole country. By means of regression curves, we also computed the distribution of average values of wind maxima for the entire country on the basis of data on wind maxima relating to the mentioned 17 instability lines from the last 10 years. The

results are presented in fig. 22. On the map, the positive and negative deviations from the national average are entered. Now if the regression line at any point of the country is used for prediction purposes, then it is correct to add the value of quadratic scattering to the average value in regions of positive deviations and to subtract the same amount in regions with negative deviations.

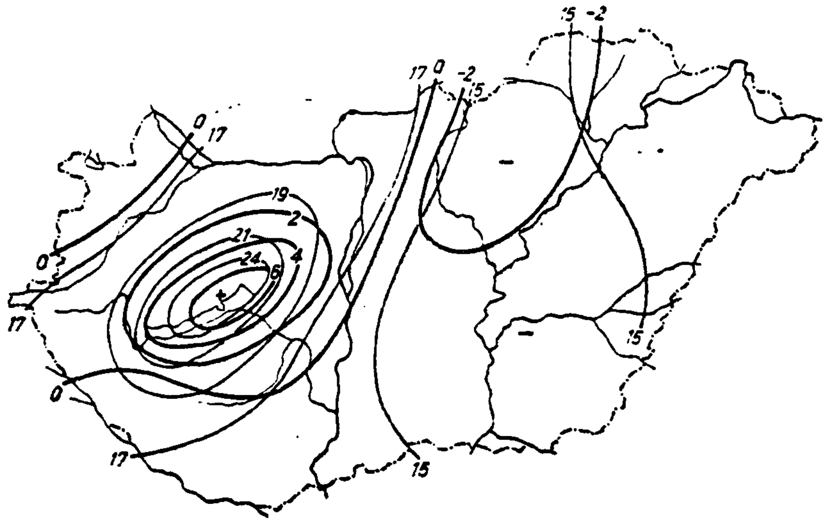


Fig. 22. Average Distribution of Maximum Wind Gusts Occurring in Connection with Instability Lines. Thin Line: Average Value; Thick line: Deviations from Average Values Computed for the Entire Country.

For convective thunderstorms, we examined the relation of the temperature jump with the max. wind gusts on the basis of 72 convective thunderstorms occurring from 1960 to 1963. The data on the temperature jump and max. wind gust was taken from the records of the Siófok observatory. For this type of thunderstorm, the choice of observation location causes no basic difficulties due to the local character of the storm. The selection of Siófok is due to its being the seat of the Storm Warning Service and to its having the needed records available. For this series of data consisting of 72 bits, the correlation coefficient has the value 0.63; the quadratic scattering is 3.5 m/s. The relation existing between the two examined parameters is illustrated by the scattering diagram in fig. 23. The equation of the solid regression line is:

$$V_{max\,k} = 7.4 + 1.9\Delta t \qquad (3.3.2)$$

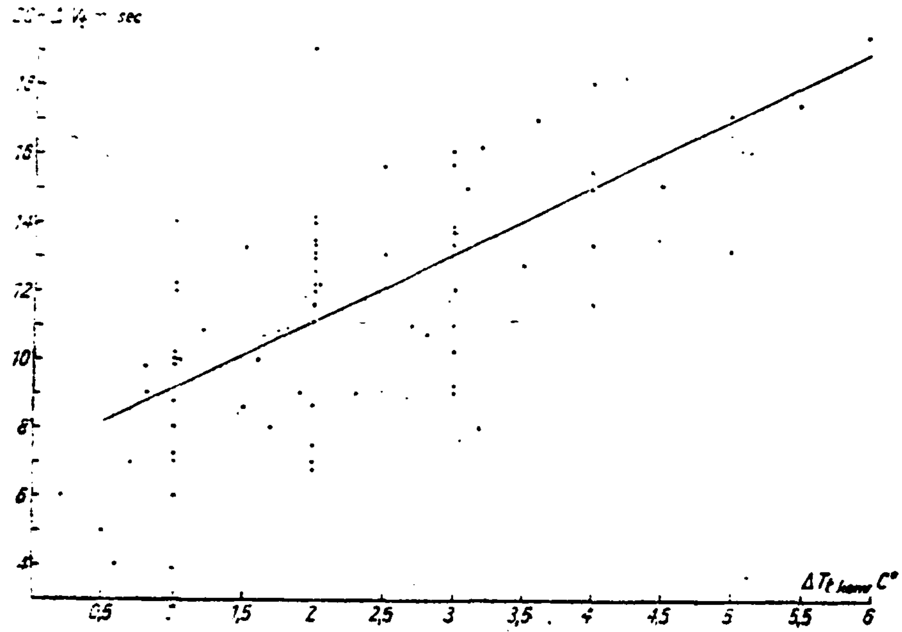


Fig. 23. Relation of the Actual Temperature Jumps with the Actual Wind Maxima in the Case of Convective Thunderstorms

This diagram and the regression equation state convincingly that the regression line valid for instability lines cannot be used to predict local thunderstorms. Whereas in the case of the instability line, a temperature jump of 1 °C corresponds to a wind gust of 14.2 m/s, in the case of convective thunderstorms, we find a value of 9.3 m/s. In a convective thunderstorm, 5-6 degree temperature jumps simply do not occur, and the corresponding wind gusts have a value of 17-18 m/s, but the same temperature jumps at instability lines cause wind gusts of 19.5-20.8 m/s.

Among the various types of thunderstorm activity there exist deviations with regard to the relations present between these important parameters. This investigation also supports the supposition that it is correct to take into account the differences present with regard to the magnitudes of spatial and chronological extent of the phenomena in the investigation of synoptic processes. A comparison of the two regression lines gives an explanation of why in practice the predicted value of max. wind speed is so unsatisfactory.

The values of the maximum wind gust and of the temperature jump for 36 cold fronts from the preceeding 10 years which were accompanied by thunderstorms, are also available through different processing. For example, the scattering diagram of the two parameters was prepared for this type of thunderstorm as well. The scattering of the dots in fig. 24 is very great, and for this reason it makes no sense to perform the above computations for this case. The result was anticipated: Since in case of fronts, the falling air masses spreading out in the course of the thunderstorm activity, are accompanied by other dynamic factors. Thus it must be stressed that even for sudden cold-front thunderstorms, the relations may not be used which can be well-defined in the case of instability lines and convective thunderstorms, and it is stated expressly that these three types of thunderstorm activity belong to three different orders of magnitude; they even have to be classified according to their sizes and their investigations must be performed separately.

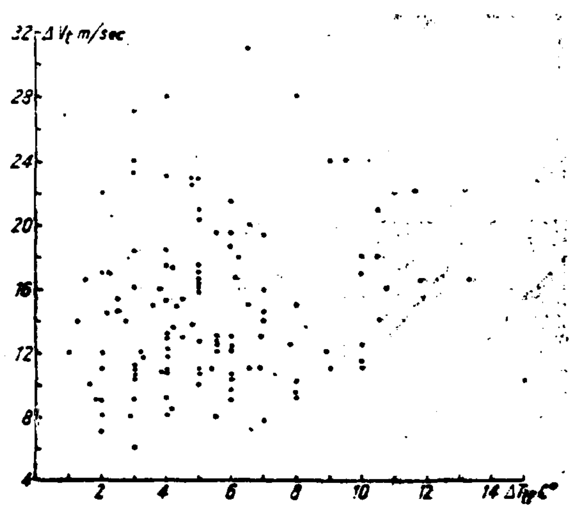


Fig. 24. Relation of the Actual Temperature Jumps with the Actual Wind Maxima in the Case of Frontal Thunderstorms

By means of the relations described here, we began to try to get an explanation of how the actual, occurring wind maxima are linked with the temperature jumps in thunderstorms. If these results are to be used in practice, then the prediction problem consists in predicting the temperature jumps in a thunderstorm.

In handling the prediction of the temperature jumps of a thunderstorm, as introduction to the thermodynamic considerations it should be pointed out how this problem can be solved. As mentioned above, Fawbush and Miller proceed on the basis of the work of Brancato assuming that the temperature distribution in the cumulonimbus cloud corresponds to the status curve of moist temperature, and that the cold air from the cumulonimbus gets down to the ground along the moist adiabate. (Unfortunately, the original paper of Brancato was not available to us and for this reason, the principle considerations are unknown for the use of the moist status curve. But it seems probable to us that this is not due to theoretical considerations, rather this selection is based on empirical facts).

Koschmieder [11] showed by theoretical means that in a thunderstorm generated inside an air mass, the temperature of the moist thermomenter remains unchanged near the ground. Ambrózy and Tänczer obtained a very good correlation between the temperature of the moist thermometer before a thunderstorm and the temperature of the dry thermometer after the storm: The correlation coefficient was 0.8.

The most basic considerations on the thermodynamic principles for the relation between the temperature jump and the wind gusts in a thunderstorm are found in the work of Faust [4]. Faust assumes that in a certain development stage of the cumulonimbus, at the back side of the cloud, a cold air mass is generated due to a melt of ice particles at the zero degree level through withdrawal of melt heat. This cold air mass moves downward, partly since it consists of colder and heavier air and partly because it still contains rain drops. A "cold air packet" forms, it falls like a unit body downward, and this is the circumstance that Faust considers important. He also assumes that there is a horizontal temperature difference of at least 5 degrees between the air in a well-developed cumulonimbus and the surrounding air. The kinetic energy of the falling air is described by the integral:

$$E = g\int_{0}^{h} \frac{\Delta T(h)}{T(h)}\, dh \tag{3.3.3}$$

whose derivative is formed by means of fig. 25.

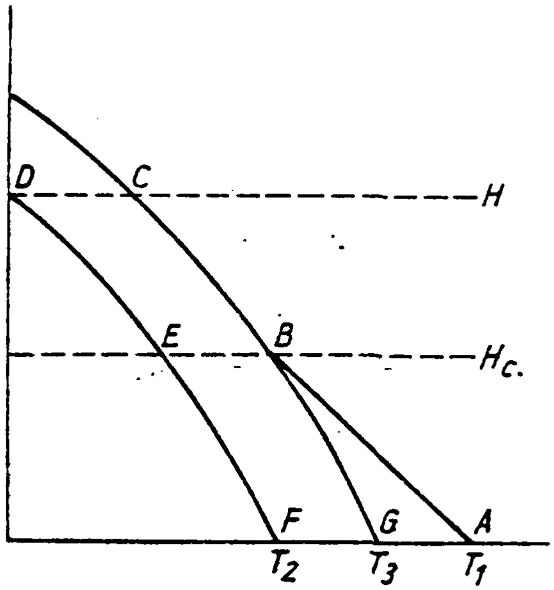


Fig. 25. Figure to Determine the Integral (3.3.3). T_1 is the Temperature before the thunderstorm, T_2 is the temperature after the Storm. ABC=status curve, DEF=moisture adiabate corresponding to temperature T_6. H_c means the condensation Level, H is the level of the 0 degree isotherm, BG is a Moisture Adiabate between the Ground and the Condensation Level *whose* Potential Temperature has value T_3 (from Faust).

At the left side of this equation is the kinetic energy E of the mass unit of air having the value $V^2/2$. On the right side, g is the acceleration of gravity, $\Delta T(h)$ is the horizontal temperature difference existing at altitude h between the falling air and the surrounding air, and T(h) is the absolute temperature of the air at altitude h. According to fig. 25, this integral can be split into three integrals in the planes DEBC, EFGB or GAB.

In a real case, the resolution of the integral for a wind gust of 32 m/s led to a value of 41 m/s. Faust improved this principle solution of the integral by using the vertical temperature gradient in the various altitude layers, and proved that:

$$\left|V_b - \frac{V_v + V_n}{2}\right| = 2R\sqrt{\frac{g}{\alpha T_1}}(T_1 - T_2), \tag{3.3.4}$$

whereby the right side of the equation is derived from integral (3.3.3). On the right side of the equation, R=0.55; α means the dry adiabatic vertical temperature gradient, T_1 and T_2 are the

temperatures before and after the thunderstorm. Faust considers his conception to be justified since a final result was attained which is similar to the empirical formula $(2\Delta T)$.

In connection with the presented equation, it should be mentioned that Faust is not speaking of a practical use of these formulas, but of how an expected temperature decrease can be predicted. Faust also is not talking about how an air particle moving down from the 0 degree level should be defined, it is only mentioned that we are dealing with the moist adiabate which corresponds to temperature T_2 of fig. 24. It is assumed that "for strongly developed Cb, horizontal temperature differences of the cold air body of up to 5 degrees compared to its environ--at least on the front side of the Cb--have to be assumed." If we were to accept this conception at face value, then we would obtain values for the temperature jump which are much higher than the actual values after applying our prediction scheme.

Findings which differ from the conception of Brancato and Faust are encountered in the work of other authors, in which the relation between temperature jump and wind maximum is examined not in a special way, but by mesosynoptic research connected with this question. In the classical work of Fujita [9] it was proven that in the case of strongly developed instability lines there is no high-altitude cold air mass, and in addition, that the so-called thunderstorm high and the cold air mass causing it, are present only up to an elevation of 5000 feet. The cooling does not occur at high altitude, but below the cloud base as a result of condensation in a dry environ. Consequently, the higher the cloud base and the drier the air underneath, the stronger is the cooling produced by condensation. This conception is expressed in the work of Pedgley [10] in which it is proven that the air pressure jump in a thunderstorm is proportional to the square of the height of the cloud base, presuming that the falling cold air is produced by the heat loss on condensation and that the moist-potential temperature is the same at all heights. This agrees essentially with the law of Koschmieder, where the temperature of the moist thermometer in the course of a thunderstorm

produced within an air mass, remains unchanged [11]. Blahotra [12] also found in an investigation of instability lines occurring in Delhi that large pressure jumps occur in a dry and warm air, i.e. at the time of the pre- and post-monsoon thunderstorms, but in the time of the Southwest monsoon, the jumps in pressure and temperature are small since the moisture content is high and the cloud base is low. In Delhi, the best month for the formation of an instability line is May; at this time the temperature gradient has a nearly dry-adiabatic value up to the level of the cloud base, and the cloud base is often at an altitude of 4 km.

The absence of a cold air mass forming at high altitude was also noticed in the chronological cross-sections of the instability lines. In the case of the observed, almost strongest instability line of 13 July 1961, the absence of a cold air mass at high altitude was also verified. Figure 26 shows a vertical, chronological cross-section for the day in question, which was prepared on the basis of the Budapest ascents. From the figure we see that the cold air assumes a place below 2 km, farther up is the undisturbed temperature field. Due to these considerations, we can state that the assumption of a cold air mass forming within the thundercloud is not convincing and this can give the physical foundation for the empirical fact that the use of the moist status curve in predicting the thunderstorm temperature jumps gives an unsatisfactory result.

The circumstances mentioned here incited us to attack the question of a possible solution to predicting temperature jumps in a thunderstorm by another direction.

The question of the temperature jump in a thunderstorm at an instability line was examined in three different ways:

1. It was assumed that the air particles in the course of the temperature jump in a thunderstorm take on a temperature corresponding to the potential temperature of the moist temperature 0° (method of Brancato). This value of the temperature jump will be called ΔT_1 below.

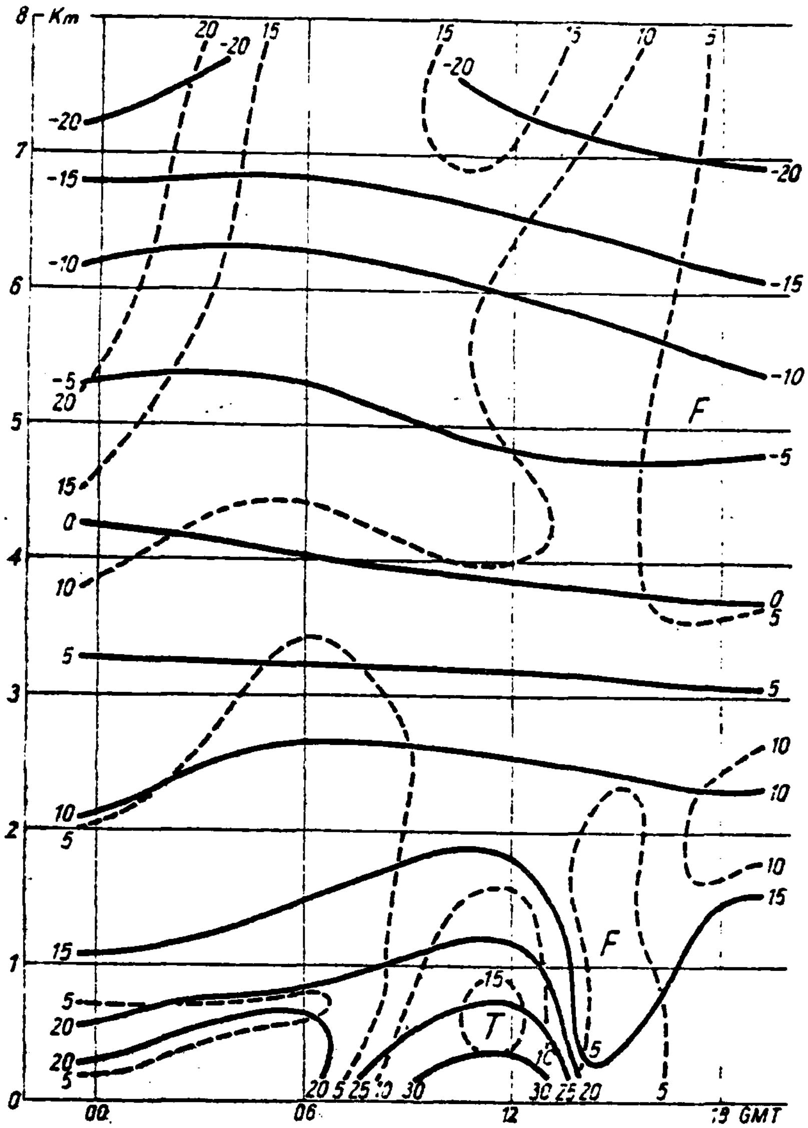


Fig. 26. Chronological, Vertical Section over Budapest, 13 July 1961 from 00 to 18 hours. The Constant Lines are Isotherms, the Dashed Lines contain the Values of the Saturation Deficit

2. It was presumed that the air particles assume that temperature corresponding to the moist-potential temperature of the 0°-point of the status curve (Method of Ambrózy and Tänczer). The value computed in this manner for the temperature jump will be called ΔT_0.

3. Finally, instead of the moist temperature before the thunderstorm, the moist-potential temperature of the condensation level is considered--a value which nearly coincides with the temperature of

the moist thermometer near the ground before the thunderstorm. The two values are not quite identical, since the value of the second quantity depends on how the condensation level is defined. This temperature is specified in the Emagramm such that the moist-potential temperature of the condensation level is used. This value then only agrees with the near-ground temperature when the condensation level is determined from the thaw point near the ground by the method of the elementary air particle. With this method it is presumed that an air particle which gets to the ground with the precipitation from the cloud, assumes the temperature of the condensation level and further cooling proceeds by itself due to the condensation occurring below the base of the cloud. The temperature difference between the temperature before the storm and that computed in the described manner, will be called ΔT_k below.

The numerical values of the temperature jumps ΔT_n, ΔT_e and ΔT_k were determined from the Budapest ascents and were then compared with the actual temperature jumps in Budapest and Siófok (ΔT_t), by using an observation series composed of 32 links. The correlation coefficient between ΔT_n and ΔT_t resulted as 0.55, the quadratic scattering was 3.7°. In order to illustrate the relation between the actual and computed values of the temperature jump, the regression equation of the two variables was determined:

$$\Delta T_n = 5{,}5 + 0{,}56 \Delta T_t. \qquad (3.3.5)$$

where $0.56\, \Delta T_t$ is the deviation from the average value of the actual temperature jump. The scattering diagram of the variables is presented in fig. 27, and the regression line computed according to the above formula is also included. If a comparison is made with the lines of slope of 45°--which represents a complete agreement--then we see that up to a temperature jump of 12° one obtains much higher values than the actual, observed values. The deviation is quite considerable for the small temperature jumps.

The correlation coefficient between the parameters ΔT_e and ΔT_t became 0.63; the value of the quadratic scattering was 3.3°; the regression equation between the computed and actual values of the temperature drop runs:

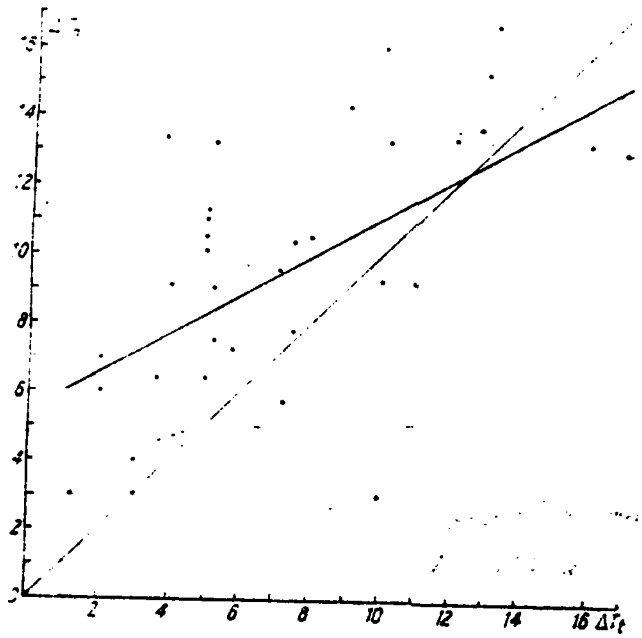


Fig. 27: Relation of the Actual Temperature Jump ΔT_t with the Computed Values of the Temperature Jump ΔT_s in the Case of an Instability Line. The Bold Line is the Regression Line, the Thin Line is a Line with a Slope of 45° which Would Correspond to 100% Agreement

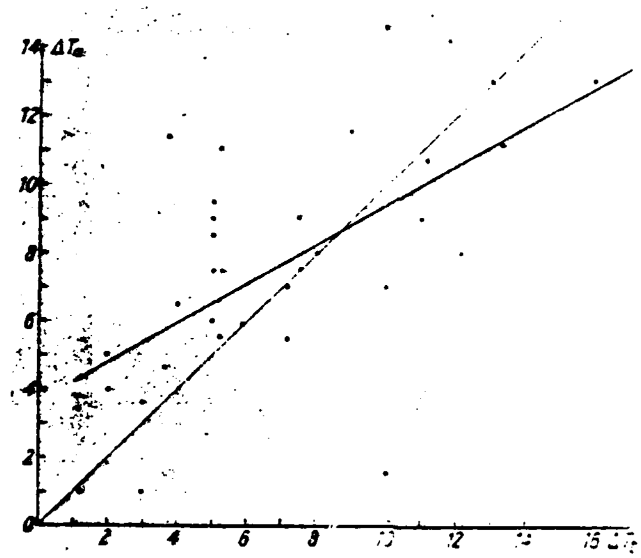


Fig. 28. Relation of the Actual Temperature Jump ΔT_t with the Computed Values of the Temperature Jump ΔT_s in the Case of an Instability Line. Presentation as in fig. 27.

$$\Delta T_e = 3{,}7 + 0{,}56 \Delta T_i. \qquad (3.3.6)$$

The scattering diagram of the two variables is shown in fig. 28. It is notable that by using this method, we obtain values having the property that they are too high below a point which nearly coincides with the halving point of the curve, and that above this point, they are too low.

Now if we consider the relations between the quantities ΔT_k and ΔT_i, then we find that they are characterized by a correlation coefficient of 0.81, and the value of the quadratic scattering is 3.1°. The regression equation runs:

$$\Delta T_k = 1{,}0 + 0{,}71 \Delta T_i.$$

The scattering diagram of the two variables and the regression line are shown in fig. 29. These results are convincing that with the moist adiabate proceeding from the condensation level, one will have the best approximation of the actual value of the temperature jump. This result agrees with the considerations of Ambrózy and Tänczer on the near-ground values of moist temperature before a thunderstorm. Thus it is expedient to compute the temperature jumps at an instability line by the last-named method, since here the scattering is smallest and the correlation coefficient is greatest. In the practical prediction of the temperature jump, the map presented in fig. 30 proved advantageous; here we find the average values of temperature jumps from 17 instability lines from the last 10 years, or the deviations from the average values for the entire country. In regions with a positive deviation, add the value of quadratic scattering to the absolute value; in regions with a negative deviation, subtract this value.

For converctive thunderstorms, likewise the quantities ΔT_n, ΔT_a, ΔT_k or ΔT_k are computed in the above manner. Here, ΔT_k is a new quantity which was not included in the previous investigation. It means a temperature jump which is defined so that the condensation level is not determined on the basis of the thaw point obtained by the ascent, rather the thaw point existing at the same time in Siófok (at the point of the convective thunderstorm) is used. For

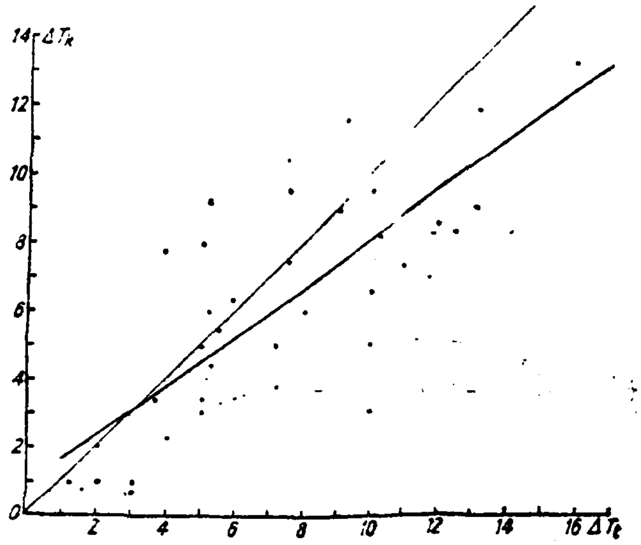


Fig. 29. Relation of the Actual Temperature Jump ΔT_t with the computed Values ΔT_k in the Case of an Instability Line. Presentation as in Fig. 27.

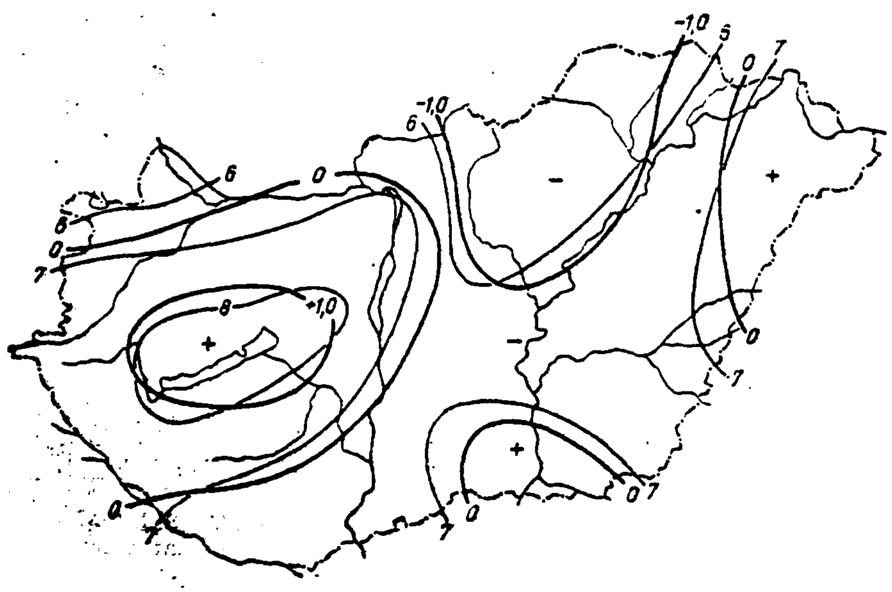


Fig. 30. Average Distribution of the Temperature Jumps Occurring at an Instability Line. Thin Line: Average Values; Bold Line: Deviations from the Average Values Related to the Entire Country.

this calculation, the ascents taken in the time before the thunderstorm were used: for morning storms and thunderstorms in early afternoon, ascents from 00 or 06 hours, and for the afternoon and early evening storms, ascents from 12 hours, and for the night

thunderstorms, the ascents from 18 hours.

In figs. 31 and 32, the scattering diagrams for ΔT_t and ΔT_n or for ΔT_t and ΔT_a are presented. As we see, the scattering of these variables is very great and the quantities ΔT_n or ΔT_a are always of a higher value than the jumps actually occurring. Consequently, the correlation coefficients also correspond to a loose relationship, since the correlation coefficient for ΔT_n has a value of 0.18 and that for ΔT_a the value 0.30. For this point-cloud, no regression curve can be drawn due to the strong scattering, since even for a linear regression for the quantity ΔT_t values are no longer obtained. Consequently, we find that this method and the hypothesis linked with it must be rejected in the case of a convective thunderstorm.

It is another situation when the actual temperature jump is compared with the moist-potential temperature of the condensation level. In figs. 33 and 34, the scattering diagrams for ΔT_t and ΔT_k or ΔT_k are shown; these speak for a good linear relation. Consequently, the correlation coefficients have a favorable value, namely, 0.70 for ΔT_k and 0.86 for $\Delta T_k'$. The regression lines entered in the scattering diagram, have the following equations: for ΔT_k

$$\Delta T_k = 1{,}0 + 1{,}1 \Delta T_t \tag{3.3.7}$$

and for $\Delta T_k'$ we have:

$$\Delta T_k' = 1{,}0 + 0{,}93 \Delta T_t \tag{3.3.8}$$

If these lines are compared with the thin line of 45° slope, then it is striking that in the case of ΔT_k, only small deviations result, and in the direction of greater values, the agreement is even better. Also, the scattering of these two parameters is less; the scattering in the case of ΔT_k has a value of 2.4^{0}; for $\Delta T_k'$ the value is 1.7°.

If we look at the methods by which the temperature jumps can be described for the two types of thunderstorms, then we see that in both cases, a significant improvement in results can be achieved by identifying the temperature occurring after the storm with the moist-potential temperature in the condensation level. In the case of convective thunderstorms, this approximation is even better

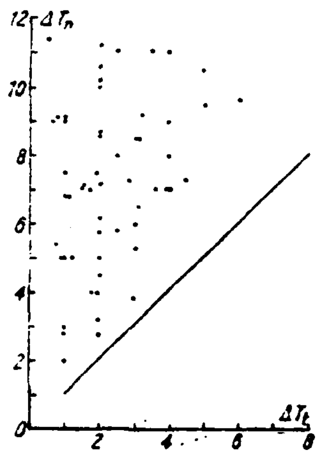


Fig. 31. Relation of the Actual Temperature Jump ΔT_n with the Computed Values of the Temperature Jump ΔT_t in the Case of a Convective Thunderstorm. Presentation as in fig. 27.

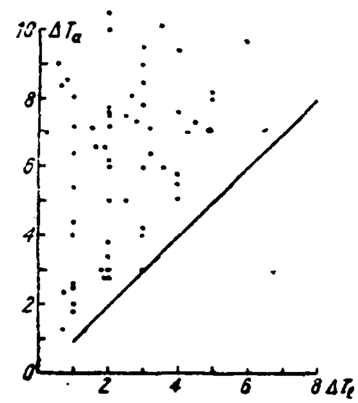


Fig. 32. Relation of the Actual Temperature Jump ΔT_a with the Computed Values of the Temperature Jump ΔT_t in the Case of a Convective Thunderstorm. Presentation as in fig. 27.

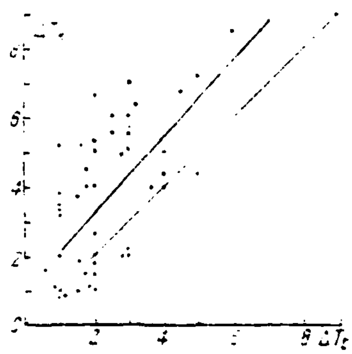


Fig. 33. Relation of the Actual Temperature Jump ΔT_l with the Computed Values of the Temperature Jump ΔT_t in the Case of a Convective Thunderstorm. Presentation as in fig. 27

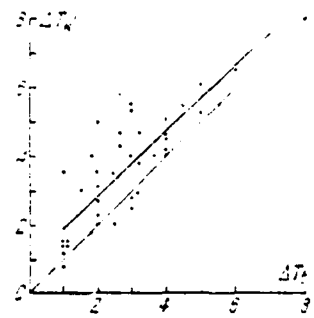


Fig. 34. Relation of the Actual Temperature Jump ΔT_l with the Computed Values of the Temperature Jump $\Delta T_t'$ in the Case of a Convective Thunderstorm. Presentation as in fig. 27

than for an instability line. For convective thunderstorms one hardly finds any deviation of a degree between the actual temperature jump and the jump computed from the local humidity. For instability lines, especially in the direction of higher values, the computed value of the jump is much too small, since here, as presumed, the heat loss through condensation plays a role. Through the obtained curves, our presumption is supported, whereby the temperature after the storm is proportional to the moist-potential temperature of the condensation level. In the case of convective thunderstorms, when

examined parameters is illustrated by the scattering diagram in fig. 23. The equation of the solid regression line is:

the cloud base is relatively low and the air beneath the cloud base is moist, there is no heat loss through condensation, or this loss is small, and consequently, the temperature after the storm is very near the moist-potential temperature of the condensation level. Conversely, in the instability lines where the cloud base is generally high and the air beneath the base dry, high values of condensation cooling occur and consequently, the temperature after the storm is lower than the moist-potential temperature in the condensation level. According to our investigations, the condensation level for instability lines is always above 2000 m, whereas for 62 convective thunderstorms, from the thaw points measured in Budapest-Lorinc, an average height of the condensation level of 1650 m was computed, and due to the dew points in Siófok, we obtained 1300 m. With regard to the actual cloud height, we have no data from the named observation posts which is based on measurement.

Through the obtained results it is proven that the relation between temperature jump and max. wind gust in a thunderstorm has a reality for thunderstorms which belong to the suitable size group and in these cases, it can be used in weather predictions. The regression lines obtained under consideration of the above presentation, can also be used for Storm Warnings on Lake Balaton.

3.4 Wind Prediction on Lake Balaton using a Method of Litvinova (E. Titkos)

A method of predicting stormy winds on the Aral Sea was prepared by Litvinova [13] (using a paper by Gubin on wind increases in changing baric gradients). This paper was applied to Lake Balaton and we came to the conclusion that the method can be used here in weather situations similar to those shown in fig. 35, i.e. in weather situations where a wedge from the Azores anticyclone extends out to the Transdanube region. This anticyclonal wedge usually lasts a long time, and during this time, the air pressure within the wedge changes, causing a local change in the baric gradient.

The relation existing between the baric gradient and the change in wind speed, is defined by the formula:

of the phenomena in the investigation of synoptic processes. A comparison of the two regression lines gives an explanation of why in practice the predicted value of max. wind speed is so unsatisfactory.

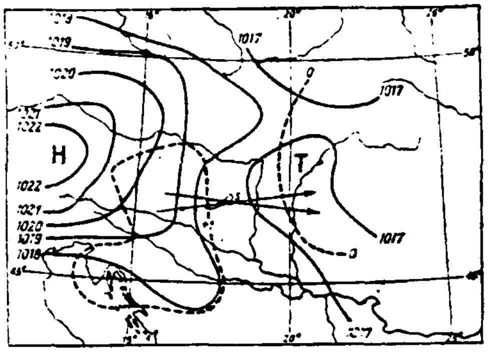


Fig. 35. Synoptic Ground Map of 7 July 1962

$$V_2 - V_1 = \frac{1}{f\varrho}\left|\nabla_H \frac{\partial p}{\partial t}\right| \cos\varphi \Delta t \qquad (3.4.1)$$

[13], whereby V is the wind speed, $\nabla_H \partial p/\partial t$ the isallobar gradient, φ is the angle formed by the baric and isallobar gradients, $\Delta t = t_2 - t_1$ is the period of validity of the prediction, f is the Coriolis parameter, and ϱ is the air density.

If only angles of 10°, 20°, etc. to 90° are assumed, and values of the isallobar gradient (as abscissa) and of the speed change (as ordinate) are plotted, then because of the above equation, we can construct the nomogram shown in fig. 36, which was prepared in the hope of being able to determine the change in wind speed in the following manner:

On an analyzed, synoptic map, two lines are drawn through the point for which the change in wind speed is to be determined; one line is parallel to the baric gradient, and the other is parallel to the isallobar gradient, whereby the directions of the affected gradients are drawn in. Then, with a protractor, measure the angle formed by the two lines; if this angle is greater than 90°, then the complementary angle (to 180°) is used. Now a distance of 125 km is marked off on the lines running parallel to the isallobar gradient, in both directions (both in the direction of larger, and in the direction of smaller, trend values), and at the two end points of the 250 km distance, the trend values of the map are read off; the absolute value of the difference of the two trend values is formed. Now this value is sought on the abscissa of the nomogram in fig. 36

are to be used in practice, then the prediction problem consists in predicting the temperature jumps in a thunderstorm.

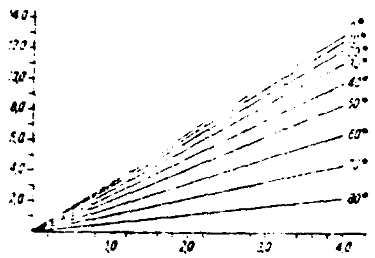


Fig. 36. Nomogram for Wind Predictions

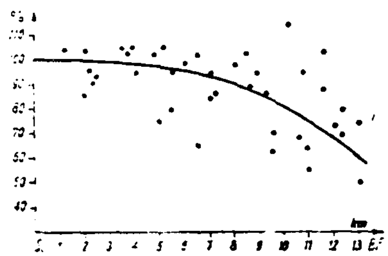


Fig. 37. Relation of Computed Wind Change with the Actual Wind Change

and from this point, move perpendicular to the particular angle value; now the change in wind speed can be read off at the ordinate. This change relates to a period of 3 hours and the change must generally be added to the instantaneous, existing wind velocity; but in cases where complementary angles were used, the change must be subtracted.

To illustrate the described method, the route of calculation of the wind velocity change in Siófok will be presented for the weather situation of fig. 35. Here, the angle between the baric and isallobar gradients is 13°. At the lines drawn parallel to the isallobar gradient, a trend value of +1.3 is read off at a distance of 125 km to the left of Siófok, and at a distance of 125 km to the right, the value is 0.0. The absolute value of the difference is 1.3. If this value is found on the horizontal axis of the nomogram, and from there if we move perpendicularly up to the value of 13°, then we obtain on the left side of the nomogram, the reading of 4.1. In our case, this amount must be added to the instantaneous value of the wind speed.

In order to test this method, in 23 cases the expected wind in Siófok was computed. In fig. 37, the relation between the computed values (horizontal axis) and the actual values (perpendicular axis) is presented. As we see, in weather situations similar to that of fig. 35, approximate information can be obtained about the wind speed change at Lake Balaton by using the described method.

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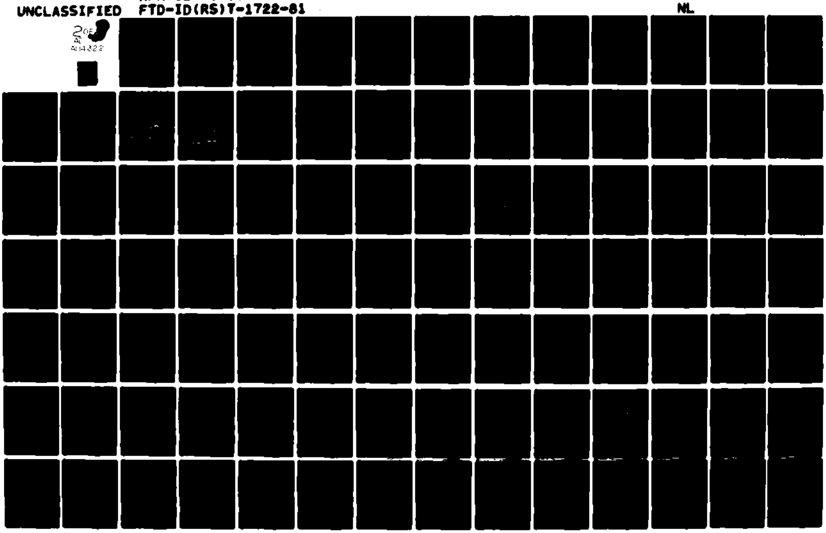


It should be mentioned that the above 23 cases in which the wind speed change was computed, come from the summer months of 1962 (June to August). In this period there was an Azores wedge over the Transdanube basin on 53 days, and in 38 cases there was a stepwise, long-lasting wind change characteristic for these weather situations, but only in 23 cases were synoptic maps available for those hours of the day in which the wind change occurred.

4. Thunderstorms in the Area of Lake Balaton

4.1 Introduction (G. Götz)

A thunderstorm consists of independent electric ground discharges of large size (lightening) taking place between different parts of the same cloud, or between the cloud and the ground; these are linked with a powerful sonic phenomenon, Thunder. The powerful voltage differences forming between various parts of the cloud or between the cloud and the ground, are generated by various, not completely understood, physical cloud processes. If the voltage gradient reaches a value of at least 30,000 Volts/cm, then an equilization occurs in the form of a lightening flash. Under the effect of the current flowing in the lightening path which has an average intensity of 20,000 to 30,000 amps, the temperature can reach 30,000 ^{o}C along this path, according to available estimates. The tremendous temperature change taking place in a fraction of a second causes a sudden expansion of the air in the lightening path (which has a cross-section of a few cm^2) which is accompanied by an explosion-like sonic phenomenon. In the direct vicinity of a lightening strike, one actually hears only a single, sharp report, the sound of distant lightening undergoes multiple reflection from the cloud and the ground and causes the long-lasting thunder roll. The length of the lightening is usually 4 to 6 km, but there are rare flashes of 10 km in length. Given this long length, the sound waves coming from the lightening path reach our ears at different times and this causes the well-known, long roll or peal of thunder which is characterized by time changes in intensity.

Nearly half of all lightening consists of flashes which strike the ground. These form a source of danger for buildings and facilities, can start forest fires or grass fires, and of course they are

a danger to man and animal when outdoors. Nevertheless, a thunderstorm on Lake Balaton poses the greatest danger to recreation not primarily by the electric discharges, but by the weather phenomena connected with it. Now it is known that the number of cases is small in which the processes inside the thundercloud progress without affecting the other weather elements. But it is much more frequent that the storm is accompanied by dangerous phenomena: The passage of a thunderstorm is often accompanied by a devastating rainfall, cloudbursts with a sudden deterioration in visibility occur and in almost all cases there is wind, frequently reaching storm strength.

From the viewpoint of the Storm Warning Service at Lake Balaton, the prediction of thunderstorm wind gusts requires the greatest attention and care. From the interior of the thundercloud there is a powerful descending air movement, simultaneous with the falling precipitation, and this air reaches the ground with an accelerated movement. Due to the presence of the solid earth, this descending air stream is forced to spread out to the sides, thus causing a horizontal flow, i.e. a wind, and near the thunderstorm border, there are sudden wind gusts. The danger of this wind consists in the fact that it begins suddenly, and there are frequent cases where the wind increase is more than 15 m/s within a few seconds.

The sudden onset of storms--as explained in section 2.2--is almost always connected with thunderstorm activity. For this reason, the Storm Warning Service cannot ignore the investigation of the mechanism of thunderstorms, their size and prediction, and where possible the reinforcement of knowledge in these areas. In this and in the following section, the results of some investigations are summarized, which were performed in the course of recent years, just to satisfy this requirement. These investigations are grouped around the following questions: Spatial distribution of thunderstorm frequency; synoptic characterization of thunderstorms on Lake Balaton; prediction primarily of thunderstorms belonging to the convective group.

4.2 Thunderstorm Frequency in the Area of Lake Balaton (G. Götz and G. Szalay)

The outline of a map of thunderstorm frequency for a given region is a rather difficult task due to the insufficiency and inaccuracy of observations. For this reason, the Climate Atlas of Hungary published in 1960, contains no thunderstorm frequency map of the country. At the same time, numerous areas of practical living, including that of a practicing synoptician, requires the precise definition of location and limits of regions of higher or lower convective activity. We intended to satisfy this demand when we began constructing the frequency map of convective thunderstorms and frontal thunderstorms in Hungary for the summer period of May to September of the years 1960 to 1964. In order to decrease the difficulties coming from the nature of the observations, the country was split into regions of 30 x 30 km, and the frequency map was constructed in such a manner that the days with thunderstorms were counted for each 900 km^2 region, without regard to whether the thunderstorm activity on a given day extended over the whole area of the region.

If we examine the thunderstorm frequency map of Hungary (fig. 38), we see that the entire country has two thunderstorm zones. Both coincide with land areas having the greatest degree of orographic features: One is the area of Mátra and Bukk mountains, and the other is in the area of Pilis and Börzsöny mountains. For these regions, the annual number of days with thunderstorms in the period from May to September is more than 45 for the individual regions. In Hungary, areas with more than 40 such storms are considered stormy. Such areas are as follows: The northern, Central Mountains, the Danube Winkel Mountains, the area around Nyíregyháza and around Szeged, the Mecsek Mountains and the Köszeger (Steinamanger) mountains, and finally, the area of the Bakony Mountains and Lake Balaton. In all these regions, the number of thunderstorms is twice as high as in the middle parts of the region between the Duna and Tisza Rivers--a region which represents the zone with the fewest number of thunderstorms.

For an investigation of the thunderstorm frequency in the vicinity of Lake Balaton, we had to begin with the recordings

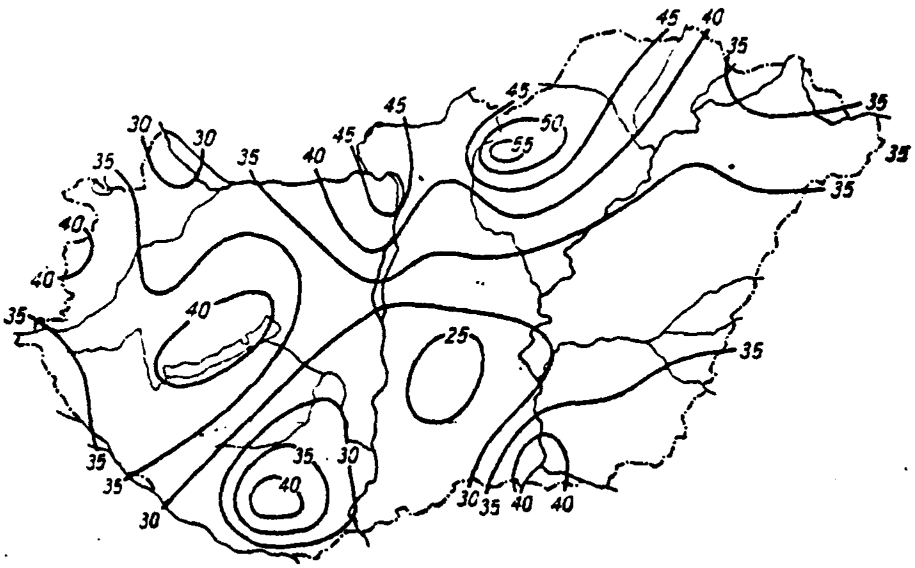


Fig. 38. Average Areal Frequency of Summer Thunderstorms in Hungary (1960-1964)

of the individual climate stations and Precipitation stations. To be sure, we were in this manner fully exposed to the inaccuracies of observations during this section of the processing, but there was no other way since the areal method used above is not suitable for describing the thunderstorm conditions of a relatively narrow region like Lake Balaton, with sufficient accuracy and thoroughness.

The average summer (May-Sept.) thunderstorm frequency for Lake Balaton and its environ was also determined on the basis of the five years of 1960-64. For purposes of this investigation, an area was selected which lies parallel to the longitudinal axis of Lake Balaton and has a surface area of approximately 130 x 70 km. Lake Balaton does not lie exactly in the middle of this region, but is shifted somewhat to the SE, since we also wanted to observe thunderstorms in the Balaton Upland. In the work, the 86 climate and precipitation stations were included in this region of 9100 km^2 (fig. 39).

The average frequency of thunderstorms in the 5 years is shown in fig. 40. As we see, those regions having more than 20 thunderstorm days are considered rich in thunderstorms. Such regions are: Zalaer Hill country, the area of Kis-Balaton, Nagyberek, Balaton upland,

Veszprémer highland, Bakony mountains, and a region bounded by the Nagycsepely-Tihany-Balatonfüzfõ-Ságvár. In the investigation of the monthly distribution of thunderstorm days, one comes to the conclusion that the month of most freqient thunderstorms in the entire Balaton region is June, with 6 to 8 thunderstorm days. The spatial distribution of thunderstorms exhibits no significant deviation among the individual months, compared to fig. 40. The thunderstorm probability related to any calender day is greatest in June, in July the values are about the same as in May, then there is a rapid decline and in September the probability is only about 6 times less than in July.

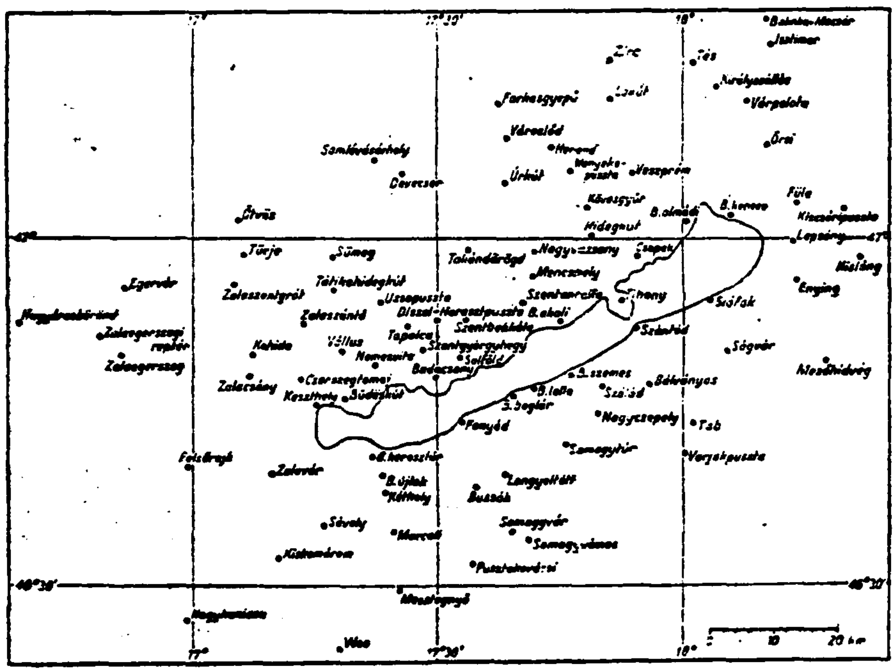


Fig. 39. The Station Network for Determining Thunderstorm Activity in the Balaton Region

It is of interest to study what kind of picture will result from a spatial distribution of thunderstorms in the direct vicinity of Lake Balaton. This was possible through provisional weather observation posts which were set up in the course of the climatological measurements of the Central Meteorologic Institute at Lake Balaton. This special network functioned in two different planes, and by using them, the thunderstorm conditions of two complete summers were

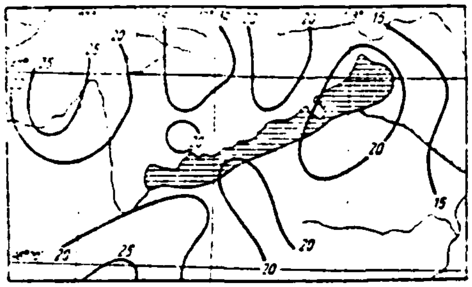


Fig. 40. Average Frequency of Summer Thunderstorms in the Area of Lake Balaton (1960-64)

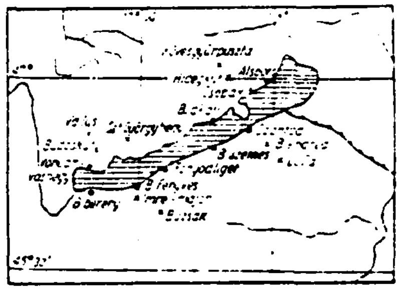


Fig. 41. Special Research Network at Lake Balaton: X = Stations of the Network in Summer 1961, ● = Stations of the Network in Summer '63.

examined. The location of the observation posts is shown in fig. 41.

In the summer of 1961, ten weather observation posts were in operation around Lake Balaton; five of them in the Southwest basin and five in the Northeast basin of the lake, and both groups were located along a line running perpendicular to the longitudinal axis in a NW-SE direction. Now this summer had the fewest number of thunderstorms of the last 10 years, consequently no conclusions of great validity can be drawn from the data. Along the line Vállus-Buzsák, it is striking that the Keszthely mountains have more thunderstorms than the environ of Nagyberek, whereas along the Kövesgyür-puszta-Lulla line, the situation is such that the Balaton upland has a large thunderstorm frequency and in the direction toward the lake, a decrease in activity is found. At the Somogyer bank, there is a recent increase in the number of thunderstorms, but in the interior of the Somogyer hill country, there is again a decrease in convective activity.

The other set of the provisional, climatic network was active in the summer of 1963; at this time the observation posts were placed on the bank itself. In contrast to 1961, this summer was exceptionally stormy. From the distribution of thunderstorms, we see that the NE basin has more thunderstorms than the SW basin, in spite of the fact that the point of greatest thunderstorm activity was in the area of Fonyód. Between the bank of Somogyer and Veszprémer, there is no significant difference; the North bank has somewhat more thunderstorms than the South bank. The smallest number was

observed in the Bay of Keszthely in this year.

Finally, we tried to classify the thunderstorms by their type of outbreak. It turned out that in the Balaton area, 62% of thunderstorms are convective ones; the others are generated by frontal effects. The fraction of convective thunderstorms is somewhat greater in the NE basin and in the mountains than in the SW basin.

4.3 Synoptics of Thunderstorms on Lake Balaton (G. Götz & G. Szalay)

The analysis of the spatial frequency distribution of thunderstorms forms only a first step in discovering the characteristics of convective activity. A complete picture of the nature of thunderstorms can only be given with the aid of observation processing for fixed observation posts operating reliably and without interruption. In the environ of Lake Balaton there are two such synoptic observation posts where observations are made by day and night. So there is hope that the primary properties of the thunderstorms moving out across the lake can be recorded. The observations posts are at Siófok and Keszthely.

For these investigations, data was selected from the observations of Siófok and Keszthely which relates to the time from May to September in the 8 years of 1957-1964. In this period there were 215 days having 274 thunderstorms in Siófok, and 185 days having 223 thunderstorms in Keszthely. The average monthly distribution is shown in tables 29 and 30. Consequently, the area of Siófok has on average more thunderstorms than that of Keszthely, and one of the reasons for this is probably that the Siófok station often treats the local-convective thunderstorm cells generated for orographic reasons over the Balaton upland, as if they were thunderstorms near the station. By the way, in the summer frequency distribution of thunderstorms, there are no significant deviations between the two stations. Thus, for the entire region of Lake Balaton, it is still true that thunderstorm activity reaches its maximum in June, July has more thunderstorms than May, in September the average number of thunderstorms drops to 2, and it can happen that in the course of

Table 29. Thunderstorm Frequency in Siófok (1957-1964)

	May	June	July	Aug.	Sept.	
Ave. no. of thunderstorm days	5.2	7.9	6.5	5.5	1.8	26.9
Ave. no. of thunderstorms	7.6	11.0	7.6	5.9	2.1	34.2
Percentage frequency of thunderstorms	22	33	22	17	6	100%

Table 30. Thunderstorm Frequency in Keszthely (1957-1964)

	May	June	July	Aug.	Sept.	
Ave. no. of thunderstorm days	4.5	7.7	5.9	3.5	1.5	23.1
Ave. no. of thunderstorms	5.6	8.6	7.6	4.4	1.6	27.8
Percentage frequency of thunderstorms	20	31	27	16	6	100%

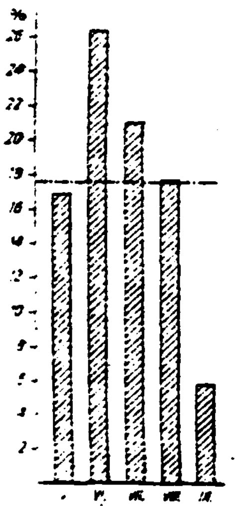


Fig. 42. Average Daily Thunderstorm Probability in the Individual Months around Lake Balaton

this pre-autumn month, no strong convective activity will occur. The average thunderstorm probability for a given calender day is 18%; the monthly values of thunderstorm probability are presented in fig. 42.

Table 31 provides information about the number of independent weather systems occurring on one thunderstorm day. It is visible that in Siófok on nearly one-quarter of the thunderstorm days, two independent thunderstorm system move over the station, and in 1% of all cases, there are four independent thunderstorm systems in the course of a single day. In Keszthely, the probability of several

Table 31. Frequency of Multiple Thunderstorm Systems forming on One Day in the Area of Lake Balaton

Number of independent systems	1	2	3	4	5
Siófok	70	24	5	1%	-
Keszthely	78	17	4	0.5	0.5%

Table 32. Percentage of Remote Thunderstorms of the Total Number of Thunderstorms in the Lake Balaton Area

	May	June	July	Aug.	Sept.	Average
Siófok	33	31	38	28	18	31%
Keszthely	20	28	12	10	8	18%

thunderstorms occurring on one day is somewhat less, but there are days having 5 independent thunderstorms.

According to the observation regulations of the synoptic stations, a distinction is made between remote or "dry" thunderstorms--symbol (☈) --where thunder is heard, but no precipitation falls at the station itself, and thunderstorms accompanied by precipitation at the station--symbol ☈. In Siófok, remote thunderstorms are found in 31% of cases, but in Keszthely, the figure is only 18%. The above explanation for the apparently greater thunderstorm activity in the Siófok area is supported by this circumstance: The convective thunderstorms forming on the NW and N-side of Siófok over the Balaton upland, never go out over the lake and are always remote thunderstorms for the Somogyer bank of the lake, but they result in an increase in the total number of thunderstorms and in an increase in the number of remote thunderstorms. This thinking is also supported by the monthly detailed processing, shown in table 32. Whereas in Keszthely, the fraction of remote storms follows the number of total storms quite closely, in Siófok, in the warmest month, July, we have the greatest number of remote storms, i.e. in a time when the conditions are best for the formation of thermal convection supported by the orography.

Similar to the investigation of stormy winds, a more complete picture of the nature of thunderstorms can be obtained by dividing them into types. There are two large classes of thunderstorm: Those within a single air mass, and frontal thunderstorms. In the former

case, the ascending air motion inside a thermally homogeneous air mass is caused by thermal convection or by forced vertical motions generated by the orography or convergence of the flow field. The thunderstorms occurring within an air mass are thus called convective thunderstorms; their spatial extent is usually of a local nature. The frontal thunderstorms occur at the boundary of two air masses, primarily at cold fronts. Since they are linked to an ascending branch of frontal air circulation, their spatial configuration is characterized by a stripwise occurrance, parallel to the front.

Of the thunderstorms occurring in Hungary, more than half are of convective type (fig. 43). An even greater fraction (65%) is found only over the northern Mittle Mts. A more detailed analysis of the spatial distribution is not the object of our investigation. But it is evident that this picture is affected by the presentation in fig. 43 which is based on the number of thunderstorm days. On a given day, both types of thunderstorm may occur. In those cases, the storm is classed as convective or frontal, depending on its intensity; this usually causes an increase in the number of frontal storms.

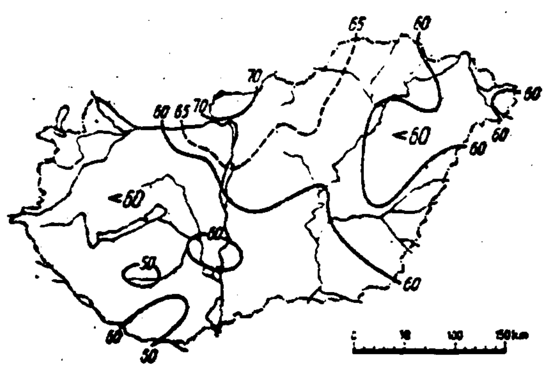


Fig. 43. The Average Fraction of Convective Thunderstorms in Hungary, Expressed in Percent of the Total Number of Thunderstorms.

Convective thunderstorms and frontal thunderstorms are only two main types of thunderstorm activity. Within these main types are numerous types of thunderstorm. Among the frontal storms, we distinguish between those belonging to fronts and frontal zones. The convective thunderstorms can be classed according to their predominate influences: Thermal thunderstorms, vergency thunderstorms and orographic thunderstorms. In addition, there are other important types of convective thunderstorms: Those occurring along a prefrontal

instability line, i.e. the squall-line thunderstorm.

In a synoptic investigation of thunderstorm activity, it is quite desirable to differentiate these different types of thunderstorm. But in order to do this, we need to have a very accurate and causal knowledge of weather processes occurring in a given area. Especially in a convective thunderstorm it is difficult to decide what importance to attach to the various dynamic factors and the updraft.

In the present treatment, no distinction was made between the various types of frontal thunderstorms (F), but the squall-line thunderstorm (S) was pulled out of the series of convective thunderstorms (K)--due to their importance. Consequently, three classes of thunderstorms are distinguished, for which the frequencies are presented in tables 33 and 34 for the areas of Siófok and Keszthely respectively.

Table 33. Percentage Frequency of Various Thunderstorm Types in Siófok

Art[1]	Mai[2]	Juni[3]	Juli[4]	Aug.	Sept.	Σ
F	25	24	38	47	53	33%
K+S	75	76	62	53	47	67%
K	75	70	54	38	47	61%
S	0	6	8	15	0	6%

Table 34. Percentage Frequency of Various Thunderstorm Types in Keszthely

Art[1]	Mai[2]	Juni[3]	Juli[4]	Aug.	Sept.	Σ
F	38	29	41	46	54	38%
K+S	62	71	59	54	46	62%
K	62	64	51	43	46	56%
S	0	7	8	11	0	6%

Key: 1-type 2-May 3-June 4-July

Convective thunderstorms are most frequent in May and June, at this time 3/4 of all thunderstorms at Lake Balaton independent of frontal activity, break forth. Toward autumn, the fraction of convective thunderstorms declines; in September, the frontal thunderstorms amount to more than half of all occurring thunderstorms. The squall-line thunderstorms are limited to the three summer months; their percentage is greatest in August. If the deviations occurring

within the Balaton region are examined, then we find that for Siófok the fraction of convective thunderstorms is greater than for Keszthely. Since the fraction of squall-line thunderstorms at both stations is essentially the same, this difference is probably due to the fact that thunderstorms forming over the Balaton uplands due to thermal-orographic factors, are observed in Siófok at a rather high frequency. But Keszthely is reached more often by cold fronts which are dissipated later in the course of their penetration inland, i.e. due to the frequently occurring Föhn effect in such cases, without precipitation, while passing the area of Siófok.

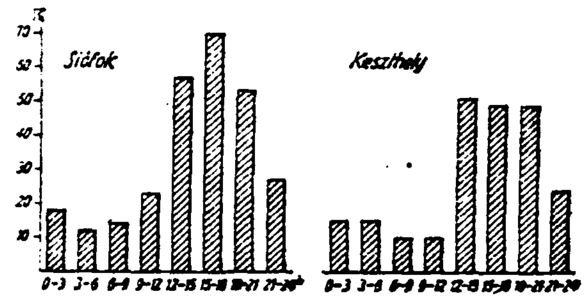


Fig. 44. Distribution of Thunderstorm Frequency by Time of Day

In the area of Lake Balaton, two-thirds of thunderstorms break out between 12 and 21 hours (fig. 44). In the environ of Keszthely, there is an almost uniform thunderstorm probability in these hours; in Siófok there is a better-developed daily profile: Between 15 and 18 hours, there is a pronounced frequency maximum. The smallest number of thunderstorms (in Siófok) occur in the time around sun-up and (in Keszthely), in the morning hours. This picture of the daily profile is explained by the daily profile of temperatures and the perpendicular equilibrium state of the air column. With regard to the fact that between the convective activity and the instable states, there is a more cohesive relation than between convective activity and ground warming, and the latter process reaches its maximum toward sundown, it is understandable that such a tendency shows up in the timing of the frequency maximum of thunderstorm activity.

If the daily profiles of the individual types of thunderstorm are examined separately, then we find (fig. 45 and 46) essentially one and the same picture. The daily profile of convective thunderstorms appears most prominent: They break out in 70% of cases in the

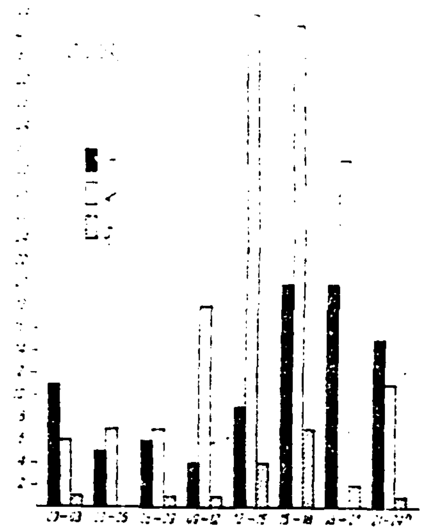


Fig. 45. Daily Profile of the Frequency of Various Thunderstorm Types in Siófok

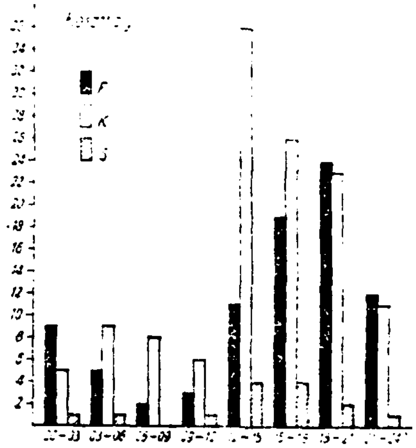


Fig. 46. Daily Profile of the Frequency of Various Thunderstorm Types in Keszthely

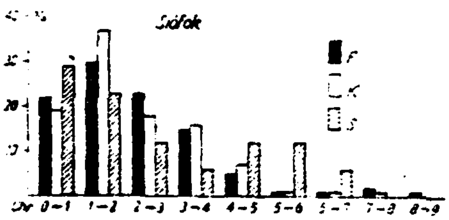


Fig. 47. Long-Term Probabilities for the Various Types of Thunderstorms in Siófok

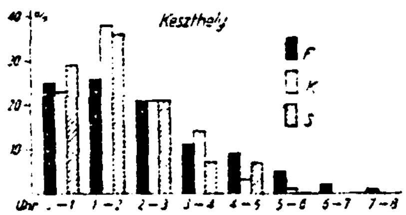


Fig. 48. Long-Term Probabilities for the Various Types of Thunderstorms in Keszthely

time between 12 and 21 hours, and the maximum frequency is between 12 and 15 hours. The frequency maximum of the squall-line thunderstorm falls between 15 and 18 hours, that of the frontal thunderstorms is between 18 and 21 hours. The frequency minimum for the outbreak of convective thunderstorms lies between 00 and 03 hours, for the outbreak of frontal thunderstorms it is in the morning and forenoon hours. In the morning hours, the squall-line thunderstorms are not able to reach the Balaton region.

The average duration of thunderstorms is 2 hours in the Balaton region. The longest duration is generally found for frontal storms having an average duration in Siófok of 2.1 hours and in Keszthely of 2.3 hours. The convective thunderstorms last 1.8 to 2.0 hours, where-

as the squall-line thunderstorms in Keszthely last an average of 1.8 hours. When the area of Siófok is reached, the duration then rises to 2.5 hours. The frequency diagrams are presented in figs. 47 and 48.

From a synoptic standpoint, the behavior of the wind is of great interest. The percentage distribution of wind structures and velocity stages occurring in connection with thunderstorms in the various types of close and remote thunderstorms are shown in tables 35 and 36 for the area of Siófok and Keszthely respectively.

Table 35. Wind Structure of Close and Remote Thunderstorms in Siófok

	F		K		S		Σ	
	ℜ	(ℜ)	ℜ	(ℜ)	ℜ	(ℜ)	ℜ	(ℜ)
1 Plötzlich ausbrechender Wind	45%	45%	65%	47%	100%	—	62%	47%
2 Stufenweise ausbrechender Wind	34	35	25	28	·	—	25	30
3 Ohne einem signifikanten Wind	·	5	3	16	·	—	2	13
4 Nur Winddrehung	7	5	6	5	·	—	6	5
5 Einstellung auf den allgemeinen Wind	14	10	1	4	·	—	5	5
6 Max. Windstoss 10 m/s	23%	22%	36%	70%	·	—	28%	58%
10—15 m/s	27	33	45	25	7%	—	36	27
15—20 m/s	30	22	18	5	12	—	21	9
20—25 m/s	18	18	1	·	31	—	9	4
25—30 m/s	2	5	·	·	25	—	3	2
≥ 30 m/s	·	·	·	·	25	—	3	·

Table 36. Wind Structure of Close and Remote Thunderstorms in Keszthely

	F		K		S		Σ	
	ℜ	(ℜ)	ℜ	(ℜ)	ℜ	(ℜ)	ℜ	(ℜ)
1 Plötzlich ausbrechender Wind	46%	14%	54%	21%	100%	—	53%	20%
2 Stufenweise ausbrechender Wind	35	29	29	33	·	—	30	32
3 Ohne einem signifikanten Wind	4	·	9	28	·	—	6	23
4 Nur Winddrehung	7	43	3	18	·	—	5	23
5 Einstellung auf den allgemeinen Wind	8	14	5	·	·	—	6	2
6 Max. Windstoss 10 m/s	22%	50%	48%	63%	·	—	34%	65%
10—15 m/s	27	33	32	26	5%	—	28	27
15—20 m/s	31	9	20	6	·	—	23	5
20—25 m/s	15	8	·	·	59	—	12	3
25—30 m/s	1	·	·	·	27	—	2	·
≥ 30 m/s	·	·	·	·	9	—	1	·

Key: 1-sudden wind outbreak 2-stepwise wind outbreak 3-without significant wind 4-wind rotation only 5-set to the general wind 6-max. gust

Now let us examine the wind structure of the various types of thunderstorm. The majority of thunderstorms, especially in the case of close thunderstorms, has a sudden onset of wind (fig. 49 and 50). The squall-line thunderstorms are accompanied without exception by the sudden onset of wind (fig. 51 and 52). A stepwise wind increase generally occurs when the thunderstorm passes by at a distance, or when the thundercloud approaches the station slowly. In such cases the wind coming from the thundercloud is initially weaker and the maximum wind gusts only occur on a closer approach of the cloud.

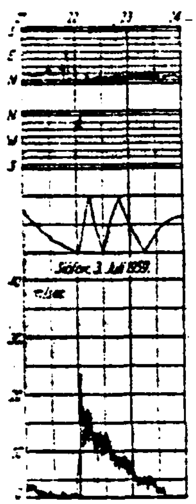


Fig. 49. Wind Recordings of a Convective Thunderstorm.

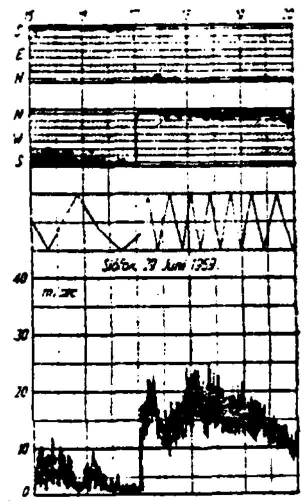
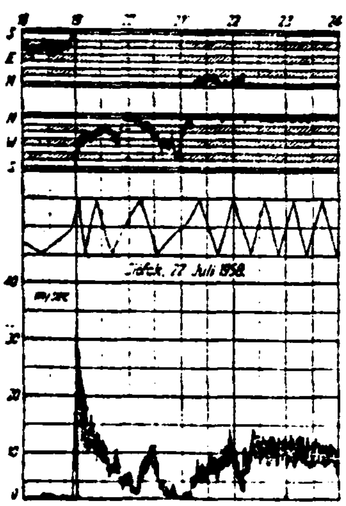


Fig. 50. Wind Recordings of a Thunderstorm Cold Front



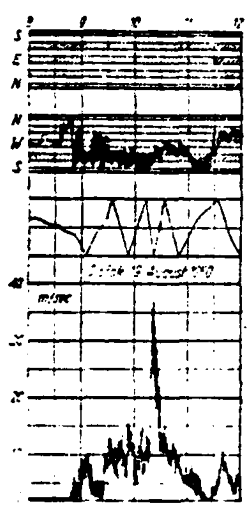


Fig. 51-52. Wind Recordings of two Squall-Lines

It also happens in the case of remote thunderstorms, or when the midpoint of the thunderstorm focus does not move right over the station, that the passage of the thunderstorm will appear on the recording chart only as a wind rotation without a characteristic change in wind velocity. One interesting case occurs when a strong wind is present at the station independently of the thunderstorm activity. If in this case the thundercloud is located in the wind direction toward the station, then the wind from the thundercloud (thunderstorm gust) is superimposed on the general wind, and causes clearly identifiable increases in wind gusts. But if the thundercloud is not moving in the direction of the ground wind, then small wind rotations can occur. In cases where the thundercloud is moving in nearly the opposite direction to the background wind, the wind from the thunderstorm causes a reduction in wind gusts. In the case of a well-formed thundercloud, the so-called emitted wind is able to suppress a ground flow of storm intensity and it can even happen that the wind will reverse for a short time in the direction of the thundercloud (fig. 53 and 54).

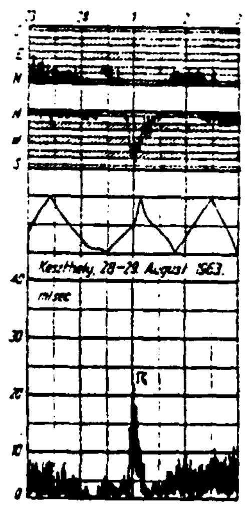


Fig. 53. Wind Recordings of a Thunderstorm activity where the Emitted Wind from the Storm has a Direction Opposite to the Direction of the Wind produced by the Baric Gradient

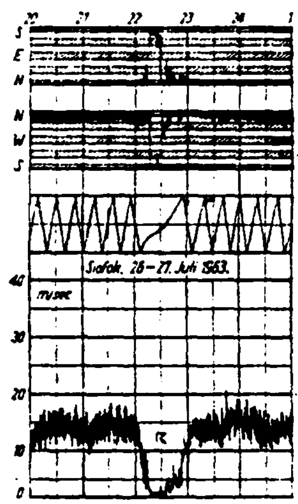


Fig. 54. Wind Recordings of a Thunderstorm Activity where the Emitted Wind from the Storm has a Direction Opposite to the Direction of the Windstorm Produced by the Baric Gradient.

Sometimes the thunderstorm is not accompanied by wind. In such cases, we are dealing either with a remote storm with an emitted wind which does not reach the station, or with a very weakly developed thundercloud having very little precipitation and in which only sporadic electric discharges occur.

The strongest wind gusts are generated in the course of the squall-line thunderstorms. In such cases, one must usually expect wind gusts exceeding 20 m/s. Also, the remote thunderstorms have weaker winds than close storms, and this seems self-evident from the above discussion. For frontal thunderstorms, the wind usually reaches storm strength, but velocities of 20 m/s occur only in 20% of all cases. Likewise, in 20% of all cases, convective thunderstorms generated over the lake, are connected with a storm strength wind. Since more than half these storms are accompanied by a sudden wind increase, their prediction is exceptionally important; but at this same time this is a very difficult task. For the Storm Warning Service at Lake Balaton, numerous investigations have been conducted in this direction; these will be reported in section 5.

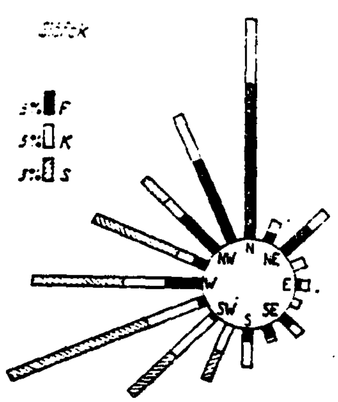


Fig. 55. Percentage Frequencies of the Wind Directions for Different Types of Thunderstorms in Siófok

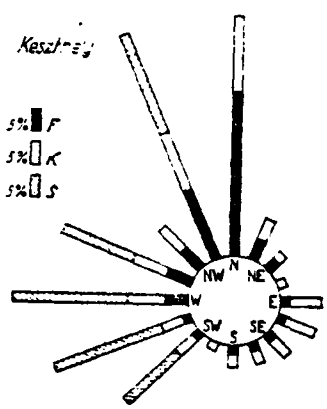


Fig. 56. Percentage Frequencies of the Wind Directions for Different Types of Thunderstorms in Keszthely

Finally, the direction distribution of winds accompanying thunderstorms will receive short notice. The precentage frequency of wind directions for the various types of thunderstorm is shown in fig. 55 and 56. Of course, this distribution contains nothing esentially new compared to that in section 2.2. Highly favored

wind directions show up for storms generated at a squall-line thunderstrom: Here, the SW direction and the W direction predominate. In the group of frontal thunderstorms, the winds blow from the NW and N sector, as a rule. However, for convective thunderstorms, there is no predominate wind direction: The wind direction is determined by the location of the thundercloud compared to the observation post and is thus not linked to any special direction.

5. Methods of Predicting Thunderstorm Activity

5.1 Introduction (G. Götz)

An important part of the work of the Storm Warning Service at Lake Balaton consists in the prediction of convective activity. The main problem is not the prediction of that thunderstorm activity which is caused by forced, rising air flows in the area of fronts. The difficulties and often even uncertainties are due to the fact that the intensity of the convective activity occurring during front-free times, even in the course of morning hours, has to be estimated on the basis of aerological, midnight observations. Namely, the emitted wind of a convective thunderstorm can often get to storm strength, and wind statistics prove that in this manner, wind gusts can occur which exceed 20 m/s by a considerable margin.

In the course of the storm warning season in 1964, which began on May 16 and ended on Sept. 20, we reported in 45 cases (i.e. 35% of all cases), the probability or presence of a thunderstorm for the Lake Balaton region in our morning predictions which are issued at 08 hours for the Water Police and which have a 12 hour validity. Thunderstorms formed within the prediction period (between 09 and 21 hours) on 45 days, i.e. just as often as predicted. In 33 cases (75% of all cases), the prediction of thunderstorm was followed by a thunderstorm; whereas in 12 cases (27% of all cases), the expected thunderstorm activity did not occur. Likewise, in 12 cases (i.e. in 9% of all issued predictions), it happened that a thunderstorm appeared in the Lake Balaton area which had not been reported in the prediction. Thus, the prediction of thunderstorms has an accuracy of 82%.

A whole series of questions arises from the data presented here. Above all, the fact that the prediction of convective activity is based on the use of modern means, is important, that is, it is not based on subjective considerations. Thus, we can speak of individual errors, but not of an exclusively subjective error. The synoptician has available at the time of prediction, the ground chart of 03 hours and the absolute topography of 850, 700 and 500 mb-surfaces from the midnight ascents (of balloons), and depending on the direction of the leading air stream, the detailed evaluation of ascents from Budapest, Szeged, Zagreb or Vienna are available on aerologic diagram paper. According to a service regulation, besides an analysis of the thermal field of the free atmosphere, the spatial distribution of the stability index of Showalter (abbreviated SSI) and of the humidity index of Lebedjeva (FI) are performed and a prediction of the status curve for the Balaton region is prepared with a 12-hour validity, and on this basis, an evaluation of the SSI and FI numbers is performed. In addition, in working out the prediction, the synoptic messages at 06 hours, the prediction maps of the German Weather Service (Offenbach a.M.) and all other maps available through facsimile transmission, are used and for the compilation of this prediction, a period of at least half an hour is specified in the service regulations.

Now are the listed aids sufficient for giving a description of expected convective activity within an acceptable error limit? The thunderstorm activity represents an unusually multifaceted process. The interaction of numerous circumstances can lead to the formation of a thunderstorm, and numerous factors are known which will suppress thunderstorms. A consideration and evaluation of factors like the instability degree, the humidity, the vergency fields of various size, the high-altitude wind and other factors, forms a confusing task and only a detailed and thorough research activity can lead to an improvement in the situation in the area of thunderstorm prediction.

Due to this composite nature of the origin of thunderstorms, the investigation methods and prognostication viewpoints can be different, as applied to this problem. The question can be studied

from a theoretical standpoint, from that of cloud physics and atmospheric electricity theory. In the areas of these two branches of science, in the course of the last decade there has been enormous progress, but unfortunately, only a part of the results found application in practical, synoptic research. It would be very useful if those parameters important for thunderstorm formation from the standpoint of cloud physics, were collected and used in synoptic meteorology. The formation of a thunderstorm can be investigated with kinematic and dynamic methods. From the baric field and the wind fields, those operators can be found which serve to describe the atmospheric vertical movements so important for thunderstorm activity. One important parameter can be found in the purely kinematic investigation of the horizontal motion field, since the wind profile in certain cases can lead to a shearing of the convective cells and in this manner negate the possibility of thunderstorm formation, even under otherwise favorable conditions. Finally, the appearance of a thunderstorm can be examined by the thermal and hygric components of the atmosphere; the importance of this factor has been reduced somewhat by recent investigations, but it is not a good idea to leave this out of consideration.

The investigations suggested from various viewpoints may naturally be performed independently in order to get a clearer idea of the role and influence of various parameters. But it should be stressed that a successful method can only be expected through the linkage of all these processes.

Among the thunderstorm investigations undertaken to support the work of the Storm Warning Service at Lake Balaton, are the kinematic, dynamic and hydrostatic research. The objective of this section consists in providing a comprehensive picture of these investigations. Naturally, the utility of these investigations is much broader than that needed by the Storm Warning Service, and the results are of interest for the other prediction services of the country.

5.2 Validity of Atmospheric Measurements of State. Prediction of the Vertical Equilibrium State (G. Götz and G. Szalay)

In the investigation of potentials for predicting the convective thunderstorm activity by hydrostatic methods, the question of importance is how much the convective activity in a given air space can be described by the measurements of radioprobe (RS) stations located within that space. The basis for posing this question is the duality which can be observed at present in the practice of synoptic meteorology. This duality consists in the following circumstances: On the one hand, it is one of the principles of synoptic meteorology which forms the most important starting point both for conventional and for numeric prediction, that the synoptic systems and the values of the atmospheric parameters undergo horizontal convection in the direction and at the speed of the leading air stream. On this basis, successful advective predicitions on the structure of the quantities of state can be made in those cases where the level of the leading stream was selected correctly and in addition, if the condition that no factors of a dynamic nature (barocline effects) take effect within the prediction period. The justification for the advective prediction can best be proven by the results obtained in the areas of barotropic, numerical predicition of the topography maps. On the other hand, the practice (irreconcilable with the above fact) exists that the expected structure of convective activity is derived almost entirely from the data of RS stations located within the given area. The prediction service, and even the workers dealing with the question of thunderstorm prediction, are satisfied that the convective activity expected within 12, 18 or 24 hours is derived from aerologic measurements taken at midnight or in the morning, even though the air column which was the object of these measurements, can have moved several hundred kilometers within the 12 hours due to the above principle. It is possible that in the course of time, an entirely different air mass can penetrate into the affected area.

This question was first touched upon in the work of [14]. Due to the midnight measurements of 12 RS stations active in the Carpathian basin, the spatial distribution of several instability indices was constructed and the thunderstorm activity in a rectangular grid of 150 x 150 km in the period from 06 to 18 hours GMT was compared with the interpolated values of the same index. This method corr-

esponds to the general practice in meteorology and we obtained e.g. for the stability index of Showalter (SSI), that the value SSI = +1 means the presence of a convective instability and a probability of a thunderstorm, occurred only in 52% of the actual thunderstorms. At the same time, the advective prediction maps of the SSI distribution were prepared on the basis of the flow field of the 700 mb surface. In this case, the results were surprisingly better: in 91% of all occurring thunderstorms, a value of the SSI of less than +1 is obtained. This fact indicates the accuracy of the advective concept.

The question of the region and period of *validity of atmospheric* measurements of state for describing the convective activity of the air column, is a problem of exceptional difficulty. The validity limits are definitely dependent on the direction and speed of the high-altitude wind, on the size of the horizontal discontinuity in macrospace, and the intensity of the barokline influences. Thus, an answer seems almost impossible for the following question: Did a thunderstorm occurring at a location in the course of the afternoon, form in such an air column as that measured at midnight at the next RS station, or in another air column (if so, which one?). But practical considerations pose the problem: Which are the RS stations to be used for a given region and weather situation for predicting convective activity?

Primarily due to experience, and partly due to results in [14], in the Storm Warning Service at Lake Balaton, a new Working Order was introduced [15]. Consequently, the aerological data from Budapest serve as a basis for thunderstorm predictions at Lake Balaton only in those cases where the streaming at the 850 and 700 mb level has a direction between 00 and 90°; however, for a SE direction, the ascents of Szeged are used, in a SW position, those of Zagreb, and for NW situations, those of Vienna are used. There is no data available on any improvement in prediction resulting from this change; but there can be no doubt that this new working order better reflects the temporal meteorological conception.

5.2.1 Characterization of Convective Activity Based on Atmospheric Measurements of State in the Affected Areas

In Hungary there are presently two RS stations in operation: Budapest-Lorinc and Szeged. To what extent is thunderstorm activity occurring on the following day described by the midnight ascents of these two stations? The details of this question were examined in the paper [16]. For comparison of measured results with thunderstorm activity, the stability index of Showalter was also used here, since this index best expresses the relation between the equilibrium state of the air column and the convective activity. This index forms an objective measuring coefficient for the thermal instability to which an air particle at the 850 mb surface is exposed.

In the investigation, only those data and parameters were used which are already available to synopticians at the time of issuing the prediction in the morning hours. The base data were taken from the period of 1 May to 31 August, 1964. This time segment contains an exceptional thunderstorm period, but there are also periods of no thunderstorms and in this manner, the parameters of the two time segments can be compared. But at the same time, this period is rather short for obtaining a final judgement on this basis; rather, it is only possible to determine whether it will pay to examine the studied relations in more detail by using a longer time period.

Among the obtained results is the fact that the ascents from Budapest and Szeged described the thunderstorms occurring at different distances from the radioprobe (RS) stations, with almost the same accuracy. With the exception of a "stagnating air column" (due to which it can be presumed that due to the instability parameter of the air column, the convective activity of the environ can be determined for a long time) there is no real physical basis from which to explain the existance of any characteristic differences. From the results, it clearly follows that the conception is generally invalid that the hope of a correct description of convection conditions should decrease with increasing distance from the RS station. When for example, on 29 June 1964 the exceptional value of the Showalter index of -5 resulted for the quilibrium state of the air column from the status measurement in Szeged, no thunderstorm occurred

in the area of the RS station, but to the Northeast of the line Salgótarján--Szolnok--Békéscsaba, a general thunderstorm activity developed. Conversely, on 26 July 1964 in Szeged, the value SSI = +5 was obtained, and on this day there was a pronounced convective activity in the immediate vicinity of the RS station (and only in this one area of the country). Now let us turn to the data of Budapest; we find that half of all thunderstorms occur at a Showalter index value of less than +1, and this applies similarly for the immediate vicinity of the Capitol as well as for the farthest point of the country.

In summary, we can say that the obtained results do not point to any close relation between SSI and thunderstorm activity which would have a physical foundation and a prognostic value. Due to the SSI, 1 to 3% better results can be obtained in thunderstorm prediction than by proceeding merely from the simple knowledge of thunderstorm frequency. If we examine the so-called thunderstorm interval (GI) which is bounded by the value SSI = -2, then we find that in Budapest, only 60% of all thunderstorms and in Szeged, only half of thunderstorms occur in this interval. Since the thunderstorm frequency in the individual area sections is generally around 50%, one could also obtain a not much worse accuracy by simply predicting the possibility of a thunderstorm on each day instead of using the Budapest or Szedeg SSI values.

The core of the question could be better stated as the relation between SSI and thunderstorm activity according to the various sky direction-sectors around the RS station. Naturally, the results do not undergo any significant change, but this provides the opportunity to perform some interesting comparisons. The difference between the SE and SW sectors of Budapest and the NW sector of Szeged was most striking--a difference which worked in favor of Budapest. Since these sectors essentially encompass the same region, the difference of ca. 10% can only be explained by the frequency distribution of the high-altitude wind. In the study period, at the 850 mb level, the air-flow directions NW, N and NE, where it was expected that convective activity in this sector would be described by ascents from Budapest, occurred in 55% of cases, whereas flows from SE and S

occurred in 13% of cases, and for this reason, the ascents from Szeged will seldom be valid for this area.

This explanation is only vaguely supported by data, since processing was conducted by neglecting flow direction, and thus such differences came into play only as a result of deviations in the frequency of the high-altitude air streams of different direction. The difference found between the NE sectors of Budapest and Szeged was also notable: for the NE sector of Budapest, the relations were 15% better than for the NE sector of Szeged. In these sectors the convective activity was described as expected, by the Budapest ascents for a flow direction of SW and W, and by the Szeged ascents for a flow of S and SW. Now the difference can be explained (at least in part) since SW and W situations occurred in 18% of cases, S and SW situations occurred only in 14% of cases.

Now if these explanations are considered valid, then those deviations occurring generally between the instability indices of the air columns of Budapest and Szeged, and the thunderstorm activity occurring in the country, can be explained. Since the Szeged measurements of state are valid for any sector of the country only in about 40% of cases, due to the location of the station and the frequency distribution of directions of the high-altitude flow, but the measurements of state in Budapest are in a relation to some region of the country in 87% of cases, then it is understandable why the instability parameters of the Budapest air column can give a 9% better description of thunderstorm activity in Hungary.

In summary, the conclusion can be drawn from the investigations that neither the Budapest nor the Szeged measurements of status are suitable in themselves for giving a description of convective activity for any sector of the region assigned to them; consequently, instability indices computed on this basis are likewise not suitable for predicting the expected thunderstorm activity. The results derived from splitting the region by directional sectors, make it seem probable that the neglect of the leading flow is responsible for the poor relations obtained, even though this circumstance can only be vaguely expressed in the data due to the method of processing.

5.2.2 Characterization of Convective Activity Based on a Prediction of the Vertical Equilibrium State by Means of the Direction of the Leading Air Flow

As regards the problem of the leading air flow, we were able to obtain a clear and final solution by comparing the results computed under consideration of directions, with the distribution described in section 5.2.1. Accordingly, the foundation of the next phase of our work is the presumption that for a given region, the convective activity of the atmosphere is best illustrated by those instability parameters computed from the measurements of state from one such RS station which lies in the direction of the leading air flow proceeding from the region in question. Namely, the advective conception--at least up to a certain mass--corresponds to the conception that the stability and instability threshold move similarly as the synoptic systems, and the direction and speed of this translation is defined by the high-altitude wind. Accordingly, the first task consisted in finding the direction of the leading air flow. The direction and speed of the high-altitude air flow was determined for the region of the Carpathian basin on the basis of the topography at 850 and 700 mb. Then, for all 7 RS stations active in the Carpathian basin (Budapest, Szeged, Vienna, Zagreb, Cluj, Uzhgorod and Poprad), we examined the ascents at 00 hours, under consideration of the direction of the high-altitude wind (on the basis of experiences, the high-altitude wind is found primarily at the 850 mb level) to see for what parts of the country they can be assumed as decisive for the characterization of convective activity. The range of validity was determined under consideration of the effective radii of the individual RS stations related to Hungary. The comparison of SSI values with the thunderstorm activity in the corresponding regions then proceeded in the usual manner.

The close relation between the values of the SSI computed under consideration of the high-altitude flow and the convective activity, is immediately evident, and thus the accuracy of the presumption is proven that the measurements of state performed at an RS station can be considered valid only for the region of a given sector, and the location of this sector is always dependent on the direction of the leading air flow. In this case, the thunderstorm activity can be described by the SSI values with an accuracy which is 12% better

than the accuracy of a prediction based only on general thunderstorm frequency, and in region GI, 87% of all thunderstorms occurred. This means that the SSI values predicted by means of the described principles are real and reliable aids in predicting convective thunderstorms, and the improvement in the relations of about 35% forms a clear and convincing proof that the consideration of the directon of the high-altitude flow is indespensible.

The presumption that the stability and instability threshold are advected analogously to the synoptic systems, satisfies essentially a requirement of the barotropic model of the atmosphere. In this model, there is namely, no wind change with altitude, consequently, an air column moves as an individual in the flow direction and at a speed of the flow, while retaining its instability parameters. It follows that the method used here is in the final analysis, the simplest barotropic prediction for the prognosis area, which consists in the fact that the air column is advected according to the principles of the persistence of the instability parameters. The good relations obtained in this manner correspond to the results attained by means of the barotropic model in other regions (mainly in the prediction of absolute topography of 700 and 500 mb), and provides a proof that the assumption of an auto-barotropy (and thus the exclusion of the possibility of development) represents a simplification of actual conditions in the case of a medium-term prediction, which can be considered acceptable.

5.2.3 Other Methods to Predict the Vertical Equilibrium State

The following factors cause a change in instability conditions of an air column: A change in wind with elevation (whereby air masses from the different sky directions meet over the particular point; they have different characteristics and this causes a change in the vertical structure of the air column), dynamic factors generated by vertical air movements, non-adiabatic processes etc. Among these factors, the decisive role is played by the horizontal temperature advection (according to the results of an analysis of magnitude) which proceeds in various ways in the opposing air layers according to their size and sign. Thus the question arises of what kind of results will be obtained on the relation between the instability

indices and thunderstorm activity for the case where advection of the air column does not occur on the basis of the autobarotropic principle, but rather if the temperature layering for the air space of the prediction region, is determined on the basis of the horizontal temperature advection.

This problem exists--since we are dealing in this case with the computation of the stability index of Showalter--in the prediction of the temperature field for the surfaces of 850 and 500 mb. Ambrózy [23] has examined this question in detail in recent years. The advective prediction for the temperature fields at 850, 700 and 500 mb for 12 hours was performed by Ambrózy on the basis of the speed and direction of the geostrophic wind pertinent to the isohypses of the affected surfaces. Now it is questionable whether the advection of the temperature fields proceeds everywhere with the corresponding wind field at the various heights of the troposphere, or not. Several descriptions are known where the temperature fields at the different altitudes are advected similar to the pressure field, with the leading flow, and the altitude of the leading flow is often identified with the 700 mb surface. For this reason, we performed the calculation of the SSI values with the method of Ambrózy, and with a prediction of temperature and dewpoint values of the 850 mb surface and the prediction of temperatures of the 500 mb surface by using the method where the advection route is defined on the basis of the 700 mb topography.

The calculations were performed for the months of the study period and for the Transdanube basin, and for comparison purposes, the autobarotropic translation of air columns was used for the same area. Through a comparison of the obtained results, one comes to the surprising but meteorologically not unique conclusions that the attempted development of the method in the interests of improving the relations, appears in the results as a deterioration, as a regression. The SSI values computed under consideration of the 12-hour advection exhibit a much weaker relation with thunderstorm activity than in the case where the air mass was subjected to a translation in the direction of the leading air flow without structural change,

and the SSI values were determined in this air column. With the prediction which was performed under consideration of the advection of the temperature fields, only for those relations obtained under the assumption that convective activity in a given region is determined by the data of one and the same RS station, were better results obtained.

All this shows a picture analogous to the results of attempts to improve the barotropic methods of pressure prediction by consideration of some barocline factors: By calculating dynamic influences, the hoped-for results were not obtained, and the accuracy of the prediction is thus reduced. To explain this fact, one must proceed with great caution, but in such cases it seems probable that we are dealing with an inherently correct consideration incorrectly built into the method. The barocline influences overlap the advective proceses usually in the form of small perturbations. Sometimes it happens that in a short time, an essentially dynamic change takes place within the system, or we encounter a dull and slow barocline effect. Now in all these cases, an extrapolation of the dynamic effects on the basis of the past and present situation, is a much more confusing and often even impossible task, but for barotropic procedures, the same method can be implemented with the greatest accuracy. Consequently, the incorrect extrapolation of barocline effects leads to large errors--even within the prediction period--whose magnitude is several times the error underlying the advection.

Sometimes it is assumed that success of the barotropic prediction is due to the mutual compensation of numerous barocline processes in the course of the prediction period, and in this manner the advection based on persistence principles is fulfilled with great accuracy. This compensation process can perhaps be explained that at each point in the atmosphere, an effort is underway to attain equilibrium with the surroundings. The occurring barocline processes always act to destroy the equilibrium, and thus other processes are initiated which try to regenerate the equilibrium state, but they are likewise of a barocline nature. Let us consider a time segment which is long enough so that the results of the two processes equal zero, and--if we only look at the beginning and end of the changes of state--the impression results that the system will undergo essen-

tially only barotropic changes of state. If only one of the baro-cline processes is arbitrarily stressed and described in such a manner as if it were a systematic, effective, sole barocline factor, then the postulated equilibrium situation is destroyed and thus a powerful increase in the error results.

Returning to the treated question, it seems certain that in reality the structure of an air column is subjected to constant transformation and temperature advection also plays a role in this process, especially since it is of differing magnitude at various altitudes and even changes its sign. But the question consists in whether the temperature field of the various high-altitude layers is actually advected in the manner as we have imagined. The in-stability indices react in a lively manner to a change in temperature values, and it can be presumed that perhaps more realistic results could have been achieved if we had proceeded by having the transla-tion of the temperature field in 500 mb not proceed at the full wind speed, rather--as is frequently done--by proceeding at two-thirds or half this speed. On the other hand, it is questionable in what manner the air column behaves under the effects of temperature ad-vection. If the air column is warmed in its lower layers, but with a cooling above, i.e. if the equilibrium situation is destabil-ized, then in a given case this can lead to a convection by which the stability is restored. Since the temperature advection is in general a slow process, such restoration of the equilibrium state can set in to the full extent, consequently in the time of the pre-diction there will by no means be an equilibrium situation which we counted on. The circumstances are different when a stabilization of the air column proceeds due to a temperature advection. A strong stability is able to exist for a longer time; consequently, on the basis of a prediction of the curve of state which was prepared by means of temperature advection, a more reliable statement can be made about the absence of convective activity than in the above case about the occurrance of thunderstorm activity. This assertion seems to be supported by the fact that in the course of our investigations, the case in which the expected, strong stability actually occurred was more frequent than the opposite case.

Naturally, by this investigation we get no answer to the new problem of what level is suitable for computing the temperature advection with the wind speed there. Somewhat better results could be obtained for the relation between SSI and thunderstorm activity if the temperature advection at all levels were computed with the wind speed of the 700 mb surface, since we have defined this speed as the speed of the leading air flow. This fact alone does not permit the conclusion that the temperature fields at the various altitudes of the troposphere are advected with this leading air flow, and that the advection routes which are computed from the gradient winds of the various elevations, could lead to no correct result. We are dealing here primarily with the relation between SSI and thunderstorm activity and this is a much too indirect relation to derive any reliable conclusions about temperature predictions. In addition, the SSI is a combination of temperature values from several elevations and for this reason, is unsuitable for deciding this question. The answer can only be obtained by performing a comparison of actual temperatures on the basis of a long observation period. A treatment of the problem in this sense is beyond the bounds of the present investigation.

5.3 Prediction of Thunderstorm Activity by Means of Instability Indices

5.3.1 Possibilities to Use the Instability Criterion of Similä in Thunderstorm Prediction for Hungary (I. Bodolai and E. Bodolai)

The most widely used methods of thunderstorm prediction are based on a calculation of the lability conditions in the middle and lower troposphere. One of the methods used to determine the vertical instability conditions of the atmosphere, is linked with the name of Similä and it is used presently in several meteorological Services for predicting thunderstorms.

The theory of the instability criterion of Similä is based on the work of Bjerknes and Petterssen [3, 17]. With this method it is necesary to have a knowledge of lability conditions of layers lying above and below the condensation level. The instability below the condensation level is defined by the thermal convection, by the near-ground friction convergence etc. The instability above the condensation level plays an important role in the development of a

cloud. According to Similä, in spite of the fact that the instability of the free atmosphere is also affected by dynamic factors, this instability should have a relatively large permanence and can thus be treated as a property of an air mass. Due to this circumstance, it is also possible that the advective changes for a given prediction region can be interpreted synoptically.

The cloud formation generated by convection is usually treated in research papers as an adiabatic process and it is presumed that the environ of the air rising in the course of convection, is at rest. In order to avoid this limitation, Bjerknes and Petterssen also took the compensating flows of the environ into account.

Similä derived the following formula on the basis of these realistic preconditions for determination of the thermal instability [18]:

$$\Delta T = \frac{T_{\Gamma'}^{500} - T_{500}}{T_{\Gamma}^{500} - T_{\Gamma'}^{500}} \cdot 10,$$

where the symbols have the following meaning: ΔT is numerical value characterizing the thermal instability; $T_{\Gamma'}^{500}$ that temperature assumed by the moist adiabate emanating from the condensation level at the 500 mb surface (or from the intersection of the 850 mb surface with the temperature-state curve); T_{Γ}^{500} is that temperature assumed by the dry adiabate emanated from the mentioned surface at the 500 mb surface; and T_{500} is the observed temperature at the 500 mb surface.

The convective cloud formation is greatly influenced by the vertical temperature distribution and by the humidity conditions of the air. Similä used the distribution of rel. humidity in the lower and middle troposphere to evaluate the tendency of thunderstorm formation. For this purpose, the humidity limit is determined via which thunderstorms occur, or under which no thunderstorms occur. This limit value resulted from the investigations of Similä at 60%. The influence of humidity on thunderstorm formation is given by the deviation Δu of the rel. humidity from the value 60%. If we wish to use increments of 5%, then the deviation Δu can assume values between +8 and -8, and the deviation is expresed in practice by the expression:

$$\Delta u = \frac{\bar{u}\% - 60\%}{5},$$

whereby $\bar{u}$ is the average value of the relative humidities at 850, 700 and 500 mb.

The determination of the thunderstorm probability is performed on the basis of the corresponding values of ΔT and Δu. In a paper appearing in 1955, Similä suggested 5 different combinations of these values to determine the thunderstorm probability [18].

Through a map presentation of the suggested value pairs, the instability can be interpreted spatially and under consideration of the corresponding advection, it can also be used for predictions.

The method of Similä whose foundation is under discussion here, is known in the Hungarian technical literature [19, 20]. The investigations on the applicability of this method in Hungary can be explained by the fact that the Storm Warning Service at Lake Balaton and other Alarm-Warning and Catastrophy Reporting Agencies repeatedly encounter the need for a greater accuracy in predicting instability and the thunderstorms connected with this instability. Thus, there arose a need to investigate the criterion of Similä, in addition to investigations on the other instability criteria which will be treated in section 5.3.2.

In the course of the investigations, the values of the computed criteria and the occurrence of thunderstorms, rain showers, or convective clouds in Hungary were compared. It should be noted that the categories of value pairs of temperature and humidity suggested by Similä were modified due to experiences gained in the course of information processing.

In table 37 are the criteria-pairs used by us.

The value pairs below are introduced into the investigation according to their ordinal numbers.

Table 37. Categories of Value-Pairs Used to Designate Instability

1.	$\Delta T < -8,$	$\Delta u < 0$
2.	$-8 < T < -4.$	$\Delta u < 0$
3.	$-4 < \Delta < 0,$	$\Delta u < 0$
4.	$\Delta T < -4,$	$\Delta u > 0$
5.	$-4 < \Delta T < 0$	$\Delta u > 0$
6.	$\Delta T > 0,$	$\Delta u < 0$
7.	$0 < \Delta T > 0,$	$\Delta u > 0$
8.	$\Delta T \geqq +2,$	$\Delta u < +2$
9.	$\Delta T \geqq +2,$	$\Delta u \geqq +2$

The instability criteria in question were determined in the period of 19 May to 30 Sept. 1962 on the basis of Budapest ascents at 00, 06 and 12 hours GMT. From the ascents at 00 hours, 131 criteria values were obtained, from the ascents at 06 hours, 114, and from the ascents at 12 hours, 133. The criteria values determined for the above times were first compared with the determined thunderstorm activity by using telegraphic precipitation reports, then by means of hourly synoptic observations, a comparison with the local weather conditions was performed.

The relations existing for the occurrence of thunderstorms with the individual criteria were examined by a comparison of index values computed on the basis of ascents at 06 and 12 hours GMT, with observations at 114 telegraphing precipitation stations. The values of ΔT and Δu were determined at 06 hours GMT in 114 cases, at 12 hours GMT in 133 cases. At the same time, we determined at how many of the 114 stations, a thunderstorm was observed, and this number was expressed in percent of 114, i.e. the rel. frequency for the areal occurrence of thunderstorms was computed. The rel. frequencies of the occurrence of thunderstorms as a function of the used criteria-pairs are illustrated in fig. 57 and 58. On the ordinate is the number of cases, on the abscissa, the ordinal number of the particular instability criteria-pair from table 37. From both charts we see that thunderstorms are rare in case of criteria-pairs 1 to 5. In the case of criteria 6 to 9 expressing an instability however, there is a clear increase in relative thunderstorm frequency. The diagram constructed from the two ascent schedules, also indicates that beginning from the criteria of value-pair 5 computed from the time 06 hours GMT, and from the criteria of value-pair 6 computed from the time 12 hours GMT, a more significant

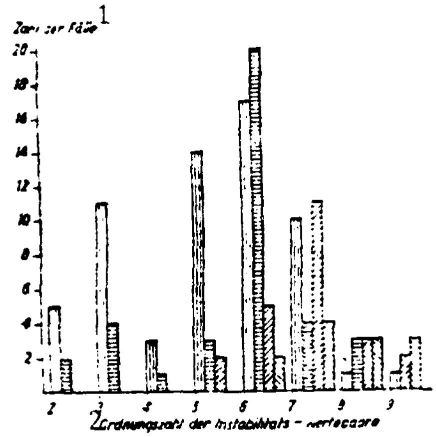


Fig. 57. Relative Frequency of Thunderstorms as a Function of the Value-pairs of Instability Computed at 06 hours GMT. The Perpendicular, dashed columns illustrate Thunderstorm-free Cases, the Horizontal, dashed Columns Illustrate Cases where 0-10% of Stations Reported Thunderstorms. Slant-dashed to Right or Left (declining) Indicates cases with 10-30% Thunderstorms or 30-50% thunderstorms Respectively

Key: 1-number of cases 2-ordinal no. of instability value-pair

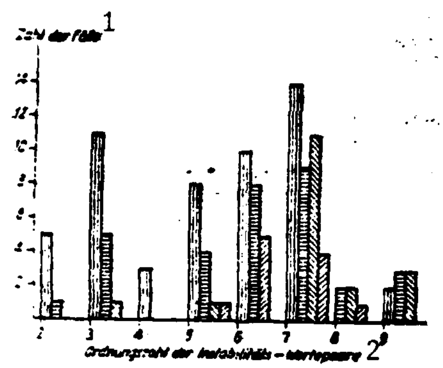


Fig. 58. Relative Frequency of Thunderstorms as a Function of the Value-pairs of Instability Computed from the 12 hours GMT Ascents. Symbols as in Fig. 57.

Key: 1-number of cases 2-ordinal no. of instability value-pair

thunderstorm activity is possible. Also, the number of cases without thunderstorms was entered into the figures. These also occur for index values expressing a stability, and for values displaying an instability, in equal numbers. Even for index value 7--which is the most frequent one--the number of cases without thunderstorms is quite large. This fact contains an indication of the limits of prognostic applicability for instability criteria and of the fact that thermal instability is only one of the basic prerequisites for the formation of convective thunderstorms.

In order to investigate the relation between instability and local weather picture, the criteria computed from the ascents at 00, 06 and 12 hours GMT were compared with the convective phenomena of the local weather picture in the corresponding 6 or 12 hours. These convective phenomena of a local nature were determined from the hourly, synoptic observations by using all observation timepoints and of the entire synoptic observation network. This finding was obtained by determining the number of stations reporting rain, thunderstorm or three special, convective cloud types, namely Cu cong, Cb calv; Cu with Sc, and Cb (i.e. the cloud types of the synoptic weather code C_L = 2, 3; 8 and 9) for the time segment of 6 or 12 hours after the ascent in question. Also, the duration of the particular phenomena was determined. In order to reduce the number of categories occurring in processing, the stations were put into groups of 3 and the time interval into 2-hour segments. Accordingly, the effectiveness of convective weather phenomena--for the sake of brevity--can be expressed by fractions in figs. 59 and 60.

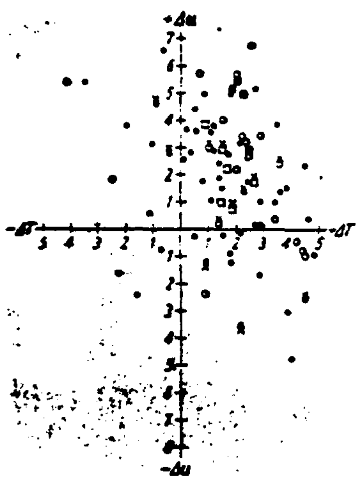


Fig. 59. Scattering Diagram of Instability Criteria in the Case of a Thunderstorm.
1/1 = ●, 1/2 = O, 2/1 = x, 2/2 = ■, 1/3 = □. An explanation of the symbols is found in the text.

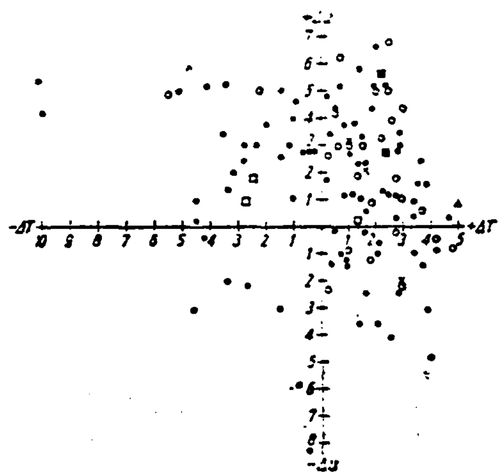


Fig. 60. Scattering Diagram of Instability Criteria in the Case of Rainshower. 1/1 = ●, 1/2 = O, 1/3 = □, 2/1 = x, 2/2 = ■, 2/3 = ▲

In these figures, the relation of number values ΔT and Δu is illustrated with the appearance of thunderstorm and rainshower. In the scattering diagrams, the cases marked by dots mean that the particular phenomenon occurred at 1 to 3 stations and lasted 1 to 2

hours. The fraction symbol for this case is 1/1, where the numerator means the above group number of the station, and the denominator means the duration of 1 to 2 hours. If the numerator is a 2, this means a number of 4 to 7 stations, and a 2 in the denominator means a period of 3 to 5 hours, and the number 3 in the denominator means a duration of 6 to 8 hours. According to fig. 59, thunderstorms occur at several stations and last longer if the quantities ΔT and Δu have large, positive values. Also, for a negative ΔT, the possibility of a thunderstorm exists, but in such cases, Δu must have a positive value. For a rainshower, the scattering diagram of quantities ΔT and Δu (fig. 60) is similar to that for thunderstorm, with the only difference, that it is more frequent that showers appear for negative ΔT and positive Δu. Thunderstorms and rainshowers can also occur for positive ΔT and negative Δu; they even occur for negative ΔT and negative Δu on an individual basis.

For a presentation of the relation existing between the individual instability criteria-pairs and the convective phenomena, table 38 and table 39 were compiled. In this presentation, daily values were computed three different times for the criteria and frequency values of the instability phenomena and used for the local weather picture. The method to produce the latter values was just discussed. In the vertical columns of tables 38 and 39, the ordinal number of the criteria, and the duration of the phenomena expressed in hourly intervals are presented. In the horizontal lines we see how many of the stations are reporting hourly rainfall, rainshower or, in the case of table 39, the three types of convective clouds.

From the two tables we see how many stations had the studied phenomenon in the case of a given criteria value for how long. It should be mentioned that in the case of the criterion of ordinal number 1, no instability phenomena occur at all. For criterion 2 there was no rainfall, but a convective cloud was observed at some places and in some cases, with the exception of Cb. For criterion 3, in four cases there was a passing thunderstorm (with a duration which falls into time interval 0 to 2 hours, and at 1 to 3 stations). It is notable that a Cb was observed in 11 cases by the same number of stations and of the same duration. In the case of the criteria

Table 38. Relation Between the Instability Criteria and the Development of Local Atmospheric Conditions (● = rain, ▽ = shower, ĸ = thunderstorm)

Kennzahl [1] des Kriteriums	Zeitdauer [2] (St.)	●	▽	ĸ	●	▽	ĸ	●	▽	ĸ	●	▽	ĸ
		1—3			4—7			8—11			12—15		
3	0—2	11	8	4	2	1	·	3	·	·	2	·	·
	3—5	5	1	·	·	·	·	·	·	·	1	·	·
	6—8	1	·	·	·	·	·	·	·	·	·	·	·
	9—11	·	·	·	·	·	·	·	·	·	·	·	·
4	0—2	14	4	·	5	·	·	2	·	·	1	·	·
	3—5	2	2	·	2	·	·	2	·	·	·	·	·
	6—8	1	·	·	1	·	·	·	·	·	·	·	·
	9—11	·	·	·	·	·	·	·	·	·	·	·	·
5	0—2	21	15	8	14	·	2	6	·	·	3	·	·
	3—5	18	2	2	2	·	·	2	·	·	1	·	·
	6—8	2	2	·	1	·	·	·	·	·	·	·	·
	9—11	·	·	·	·	·	·	·	·	·	·	·	·
6	0—2	15	17	13	4	1	3	1	·	·	1	·	·
	3—5	6	13	5	3	1	·	1	·	·	·	·	·
	6—8	2	·	·	·	·	·	·	·	·	·	·	·
	9—11	·	·	·	·	·	·	·	·	·	·	·	·
7	0—2	34	14	12	16	3	7	7	·	·	5	·	·
	3—5	27	11	9	14	1	1	5	·	·	1	·	·
	6—8	·	1	3	2	5	·	·	·	·	·	·	·
	9—11	2	·	·	·	·	·	·	·	·	·	·	·
	12—14	·	·	·	·	·	·	·	·	·	·	·	·
8	0—2	12	6	5	5	·	·	2	·	·	·	·	·
	3—5	8	7	6	1	·	·	·	·	·	·	·	·
	6—8	1	·	1	2	·	·	·	·	·	·	·	·
	9—11	·	·	·	·	·	·	·	·	·	·	·	·
	12—14	·	·	·	·	·	·	·	·	·	·	·	·
9	0—2	10	9	3	6	3	·	6	·	·	2	·	·
	3—5	10	4	3	3	·	·	2	·	·	1	·	·
	6—8	3	2	2	1	·	·	·	·	·	·	·	·
	9—11	·	·	·	1	·	·	·	·	·	·	·	·
	12—14	·	·	·	·	·	·	·	·	·	·	·	·

Key: 1-identification of criterion 2-duration (hours)

pair 4, a Cb was observed only in two cases, once lasting 2 hours and once lasting 5 hours. In table 39 in the Cb column, the number of thunderstorms is given which belong to the particular column or line of the table. With the increase in the ordinal number of criteria (beginning at criterion 5), we find in each column and in each line, an increase in frequency of occurrence of thunderstorm and rainshower, but the frequency of non-convective rainfall is exceeded only from criterion 8. The greater rainfall frequency is also enhanced since after a convective rainfall stops, there is often still some rain. If we compare table 38 with table 39, then it is striking that the relatively frequent convective clouds in many cases do not lead to formation of a thunderstorm. On the other hand, in table 39 our attention is directed to the fact that a convective

Table 39. Relation Between the Instability Criteria and Convective Clouds

Kennzahl des Kriteriums 1	Zeitdauer St.) 2	Cu cong und Cb calv	Cu + Sc	Cb	with Cb mit Cs	Cu cong und Cb calv	Cu + Sc	Cb	with Cb mit Cs	Cu cong und Cb calv	Cu + Sc	Cb	Cu cong und Cb calv	Cu + Sc	Cb
		1—3				4—7				8—11			12—15		
2	0—2	3	4	.	.	2	2	.	.	.	.	.	.	.	.
	3—5	3	5	.	.	1	.	.	.	.	1	.	.	.	.
	6—8	3	.	.	.	.	.	.	.	.	.	.	.	.	.
	9—11	.	.	.	.	.	.	.	.	.	.	.	.	.	.
3	0—2	9	13	11	4	13	7	1	.	7	.	.	4	.	.
	3—5	15	10	4	.	4	3	.	.	3	.	.	.	.	.
	6—8	7	2	1	.	1	1	.	.	.	.	.	.	.	.
	9—11	2	1	.	.	.	.	.	.	.	.	.	.	.	.
4	0—2	3	5	1	.	5	3	.	.	1	2	.	.	.	.
	3—5	2	5	1	.	2	3	.	.	.	1	.	.	.	.
	6—8	3	1	.	.	.	.	.	.	.	.	.	.	.	.
	9—11	1	.	.	.	.	.	.	.	.	.	.	.	.	.
5	0—2	4	11	9	8	12	13	1	2	6	1	.	5	.	.
	3—5	17	16	8	2	7	.	.	.	2	2	.	1	.	.
	6—8	10	4	3	.	1	1	.	.	.	.	.	.	.	.
	9—11	.	1	.	.	.	.	.	.	.	.	.	.	.	.
6	0—2	19	22	18	13	23	5	6	3	14	1	.	.	.	.
	3—5	35	26	17	5	9	3	2	.	5	.	.	.	.	.
	6—8	17	11	4	.	2	1	.	.	.	.	.	.	.	.
	9—11	.	.	1	.	.	.	.	.	.	.	.	.	.	.
7	0—2	[illegible]	32	17	12	32	18	10	7	26	1	4	8	1	.
	3—5	40	27	18	9	11	12	6	1	1	.	2	.	.	.
	6—8	17	9	4	5	5	.	.	.	.	.	.	.	.	.
	9—11	3	1	.	.	.	.	.	.	.	.	.	.	.	.
	12—14	2	.	.	.	.	.	.	.	.	.	.	.	.	.
8	0—2	6	5	8	8	6	4	3	.	5	1	2	4	.	.
	3—5	8	10	7	6	6	1	1	.	1	.	.	.	.	.
	6—8	4	3	5	1	1	.	1	.	.	.	.	.	.	.
	9—11	3	.	.	.	.	.	.	.	.	.	.	.	.	.
	12—14	1	.	.	.	.	.	.	.	.	.	.	.	.	.
9	0—2	7	7	9	3	5	7	3	.	1	1	1	.	.	.
	3—5	5	9	7	3	2	2	2	.	.	.	.	1	.	.
	6—8	7	4	1	2	2	.	1	.	.	.	.	.	.	.
	9—11	1	.	.	.	.	.	.	.	.	.	.	.	.	.
	12—14	1	.	.	.	1	.	.	.	.	.	.	.	.	.

Key: 1-identification of criterion 2-duration (hours)

cloud itself can occur rather often under stable circumstances, and this must be taken into account when preparing cloud predictions. Tables 38 and 39 give convincing proof that the thunderstorms can be predicted with sufficient certainty only in cases of criteria 8 and 9.

In order to study the instability criteria in real synoptic situations, the values of the instability criteria were determined

Table 40. Relation between the Instability Criteria and the Development of Local Atmospheric Conditions in Cases of Some Storms in the Summer of 1962. (o = rainfall, ▽ = rainshower, ☈ = thunderstorm)

1 Datum	2 Kennzahl des Kriteriums	Form	3 Häufigkeit	2 Kennzahl des Kriteriums	Form	3 Häufigkeit	2 Kennzahl des Kriteriums	Form	3 Häufigkeit
		4 der konvektiven Erscheinung			4 der konvektiven Erscheinung			4 der konvektiven Erscheinung	
20—21. Juni 1962 12, 18, 00 GMT	6	▽	1/1	7	Cb	1/2	7	☈	1/2
27. Juni 1962 00, 06, 12 GMT	7	▽	2/2	6	▽	1/1	6	▽ ☈	1/2 1/2
2. Juli 1962 06, 12 GMT	7	▽ ☈	1/1 1/1	9	▽	1/2	.	.	.
12. Juli 1962 00, 12 GMT	6	☈	1/2	9	☈	1/2	.	.	.
13. Juli 1962 12 GMT	9	▽ ☈	1/3 2/2	.	.	.	.	.	.
15. Juli 1962 00, 06, 12 GMT	7	● Cu—Sc	1/1 1/1	7	● Cu+Sc	1/2 2/2	7	▽ ☈	2/1 1/2
[illegible] Juli 1962 [illegible] 12 GMT	7	●	1/2	7	● Cu—Sc	1/2 2/2	7	● Cb	2/2 1/2
22. Juli 1962 12 GMT	6	▽ ☈	1/1 2/1	.	.	.	.	.	.
27. Juli 1962 12 GMT	6	● ☈	1/1 2/1	.	.	.	.	.	.
7. Aug. 1962 12 GMT	6	Cb Cu+Sc	1/1 1/2	.	.	.	.	.	.
15—16. Aug. 1962 12, 00 GMT	6	Cb	1/2	6	▽ ☈	1/2 1/1	.	.	.
17. Aug. 1962 12 GMT	9	▽ ☈	1/1 2/1	.	.	.	.	.	.
18. Aug. 1962 00, 06, 12 GMT	9	☈	1/1	8	▽ ☈	1/1 1/1	7	▽ ☈	1/1 1/1

Key: 1-date 2-identifier of criterion 3-frequency 4-of convective phenomenon. Juni = June; Juli = July.

for several such cases from 1962, in which rainshowers and thunderstorms occurred with the storms. In table 40 we have the values of ΔT and Δs computed from the ascents near in time to the outbreak of the studied, convective phenomena. The symbols introduced above were retained. From the table we see that the criterion of Similä—regardless of the genetic form of a convection occurrence--gave a true reflection of the instability conditions in all cases, and a higher ordinal number than 6 did not occur among the named cases.

Through the investigation presented here, it is proven that the instability indices of Similä offer a good picture of the instability state of the atmosphere. From the obtained results it is visible that the criteria values from 1 to 4 represent stability, from 5 to 7 a moderate instability, from 7 to 9 a strong instability.

In the case of criteria 5 to 7, the diagnosis of the synoptic situation should be studied with extreme care in the prediction, since for these values of the criterion, the number of cases with and without thunderstorms are equal. At values of 8 and 9, the thunderstorm can be predicted with great certainty.

5.3.2 Relation of the Various Instability Indices with the Convective Thunderstorm Acitivity (G. Götz and G. Szalay)

To predict the lability conditions of an air column, in section 5.3.1 the lability index of Similä is used. Now the question must be examined as to what differences result in the prognostic value of the various instability indices, if they are applied by the same methods to the same time period.

The posed question has been touched in the course of our earlier investigations [14]. That this problem has come up again at a recent daily conference, is due to the fact that we did intend to give new instability indices untried in Hungary, an applied test and since this question is important from the viewpoint of multi-parameter thunderstorm prediction methods which will be discussed in section 5.4. Besides the instability index of Showalter and the factor K also used in the Hungarian weather service and defined by the formula:

$$K = (T_{850} - T_{500}) + T_{d850} - (T - T_d)_{700}$$

as a third characteristic quantity, the uplift index (LI) was selected.

The LI has not previously been used in the Hungarian meteorologic service. This index was prepared in the United States by the Severe Local Storms Forecast Center in Kansas City in 1956 with the objective of determining the location and size of regions with latent instability in a real manner which can be obtained by an extrapolation of the instability regions indicated by the SSI. The factor LI is defined by the temperature difference existing between the actual temperature of the 500 mb surface and a hypothetical 500 mb temperature, which is assumed by an average air particle if it were raised from an altitude of 3000 feet above the modified ground-surface to the level of the 500 mb surface. The resulting number values have the name of "lifted index". LI forms a coefficient for the lability of air layers located above the condensation level; this value has a

negative sign for air particles which reach the 500 mb level at a temperature which is higher than the surrounding temperature. Consequently, a similarity exists between LI and SSI, with the exception of the specification of that level from which the ascent of the air particle begins, and excepting the fact that SSI is a statistical index based on observations and LI is an identifier based on a prediction.

The computation of LI occurs in the following manner. First, for the lowest air layer with a thickness of 3000 feet (thus practically a layer located below the 900 mb surface), the actual average value s_m of the specific humidity is determined by graphic means by using the method of equal surfaces. Now the maximal temperature T_{max} for the afternoon is predicted and the intersection of the isoline corresponding to the temperature T_{max} (i.e. the dry adiabate running out over the point T_{max}) is determined with the isogram of S_m. By means of this point, the lift-condensation level is determined for the average air particle of the lower layer. Now from this point, the air particle along the moist adiabate crossing the condensation level is lifted up to the 500 mb surface, where the individual temperature is read off the diagram and the lifting index can be computed on the basis of the formula:

$$LI=T_{500}-T'_{500}$$

To compute LI, a diagram can be constructed.

In the course of our investigations, the determination of LI was changed somewhat. In the United States, the index is computed from the ascents at 09 hours CST, and conclusions are drawn from this about the instability at 15 hours CST. We did not have available measurements at the appropriate times, and we had to perform the investigations on the basis of ascents at 00 hours GMT. By this method, we would have too small values for s_m. This difficulty was avoided by replacing the value s_m by the value s_o (spec. humidity on the ground). Under normal circumstances, it can be presumed that this quantity lies closer to the values of s_m valid at 10 hours GMT.

For an objective determination of the maximum temperature, the following, well-known consideration was used. On sunny days when

conditions are not disturbed by fronts and intensive temperature advection, for the temperature maximum near the ground, certain limits are set by the temperature conditions prevailing in the vicinity of the condensation level. Up to the convective condensation level (i.e. usually into the layers lying below the 850 mb surface), no vertical temperature gradient much larger than the adiabatic gradient can form, since for overheating of near-ground layers, the convection will begin immediately and in the place of the overheated air body, a cooler air will flow in from the surroundings and from above. In the case of an adiabatic value of the vertical gradient, the 1000 mb surface at negative temperatures is about 11-12 degrees warmer than the 850 mb surface, and for positive, near-ground temperatures, this difference is 13-14 degrees. Since at the 850 mb surface, the daily fluctuation of temperature occurring due to radiation is much less than near-ground, in sunny weather the temperature of this level at 00 hours GMT offers a reliable support point for predicting the temperature maximum near-ground. In the course of the study period, on those days when the radiation influences predominated, at the 850 mb surface we had from 00 hours GMT until 12 hours GMT an average temperature increase of 0.3°, and from 00 hours GMT until 18 hours GMT, an average increase of 1.4°. Thus, at the 850 mb surface from midnight until the max. temperature occurred on the ground, a temperature increase of 1° must be expected, and this means that (under consideration of the standard values of temperature differences presented above between the 850 and 1000 mb surfaces) the expected max. temperature on the ground can be computed such that one adds the following quantities to the 850 mb temperature measured at 00 hours: Below 10°, add 14°, above 10°, add 15°.

The validity of this discussion is illustrated by the scattering diagram of fig. 61, where the relation of the T_{850} values at 00 hours is illustrated with the actual, occurring max. temperature T_{max} on the particular day, for all sunny days in the study period. According to our investigations, the scattering can be decreased substantially even if the expected advection at the 850 mb surface is brought into consideration. Since the advection conditions can be computed from the midnight topographies until the time of issuing the prediction, by using the described method, objective and reliable values can be obtained for the temperature max. on sunny days.

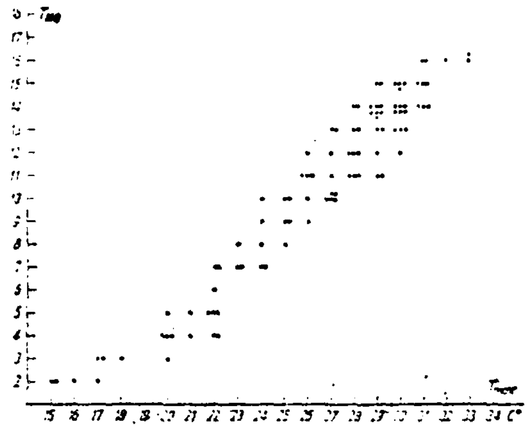


Fig. 61. Scattering Diagram of the Actual Temperature Maximum (T_{max}) and the temperature measured at 00 hours GMT at the 850 mb Surface (T_{850}), for the area of Budapest

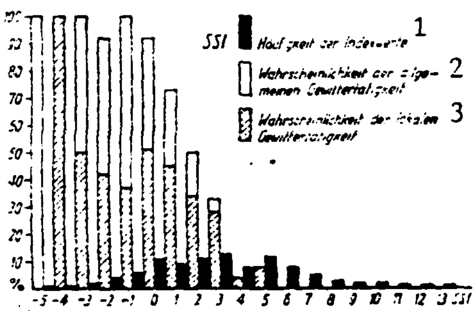


Fig. 62. Frequencies of SSI Values and the Corresponding Thunderstorm Probabilities

Key: 1-frequency of the index value
2-probability of general thunderstorm activity
3-probability of local thunderstorm activity

On the basis of a prediction of the instability indices performed by means of the autobarotropic persistence principle described in section 5.2 under consideration of the prevailing air flow, the relations of the values SSI, LI and K are presented in figs. 62-64 with the thunderstorm activity for the study period of 1 May-31 August 1964. In order to explain the details of the interrelations of the indices with convective activity, in the course of this investigation, a distinction is made between cases of occurrence of local thunderstorms and cases with a general and extended thunderstorm activity. Since for LI no concrete data is available on the width of the thunderstorm interval (GI), this interval was specified by the definition in the paper [16] in the interest of an objective comparison: For all three indices, those values are considered to be the limits of GI, whose cumulative frequency is equal to the

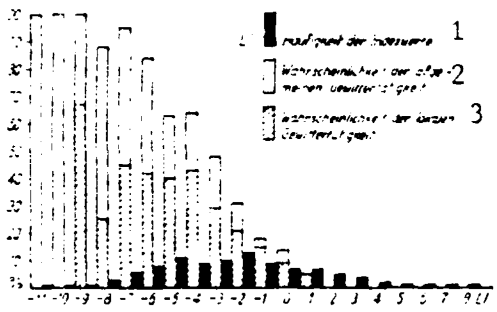


Fig. 63. Frequencies of LI-Values and the Corresponding Thunderstorm Probabilities

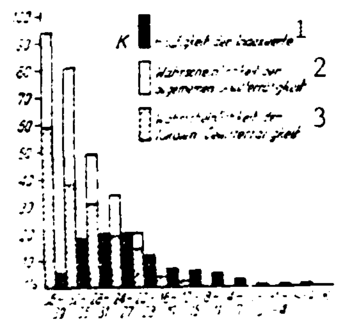


Fig. 64. Frequencies of Values of Factor K and the Corresponding Thunderstorm Probabilities

Key: 1-frequency of the index value; 2-probability of general thunderstorm activity; 3-probability of local thunderstorm activity

thunderstorm frequency. With consideration that in the present study we are dealing with one and the same time segment and with the application of the same methods, such a procedure is permitted, since the thunderstorm frequency has the same value in all three cases.

For the limits of GI, we obtained values SSI = +1.6, LI = -3.9 or K = 28.6. Within this interval, SSI described 82%, and LI and K described 72% of all thunderstorms, which proves that all three indices have a real relation with the convective thunderstorm frequency, and that they have a prognostic value. At SSI, 92% of all thunderstorm activity falls into the GI, for LI the figure is 79%, for the factor K, only 76%. In the case of SSI and LI, the thunderstorm probability is 100% at the maximum index values, and even the probability of a general thunderstorm activity is high in this case, whereas in a decrease in values of the instability indices, the probability for a general thunderstorm activity declines with respect to the local thunderstrom activity. This information indicates that prognostic conclusions affecting the extent of thunderstorm activity can be drawn on the basis of the values of this instability index if a certain caution is applied.

No doubt the closest relation with convective activity exists for the index SSI, and the relation is weakest for the factor K.

The higher prognostic value of LI compared to K shows up in the verification characteristics and in the description of thunderstorm probability, so that in the final analysis, it seems advisable to proceed from the values of the lifting index in the treatment of such problems.

In order to be able to find other, less striking differences with regard to the prognostic value of the indices, and in order to examine the reality of thunderstorm probabilities corresponding to the individual index values, the reader is referred to figs. 62-64.

This segment of the study had to decide the question of why weaker relations resulted between the LI values and the thunderstorm activity, than between the SSI values and thunderstorm activity. The lifting index was worked out with the goal of correcting some problems with the SSI; here LI also contains prognostic elements; thus one could justifiably expect that these viewpoints would also be expressed in the results.

The answer to this question is quite simple. At any rate, the problem can be returned to the examination of the starting data on the air particle. The point is illustrated here by the starting temperature (predicted value of T_{max}). The objective method described at the beginning of this section for predicting T_{max} was used without any reservations, i.e. the accuracy was ignored. By this process we tried to preserve the objective character of the investigation, since this is an important factor in operative predictions. But in certain cases, even at the time of preparing the prediction, there is no doubt that unreal temperature values result from this method; a fact which can be corrected by knowledge of the synoptic situation and by using other prognostic methods. Together with the modified value of T_{max}, the value of LI undergoes a change, and it can be presumed that in the majority of cases, this change may lead to a more correct description of convective activity.

The other difficult point in computation of LI consists in the selection of the dewpoint T_d, the knowledge of which is necessary to determine the convective condensation level. If we were to follow

the original definition, then the average state of the lowest air layer of 100 mb thickness would have to be used for the time 10 hours GMT. Conversely, we only have available a sufficient amount of data for the time 00 hours GMT. We tried--as mentioned--to go around this difficulty by replacing the average value of T_d by the ground value at 00 hours GMT, presuming that this value would give a better approximation of the future average value, since it is larger.

But this presumption of the daily profile of spec. humidity did not prove correct--as follows from subsequent studies. On days when no replacement of air mass occurred, there is an increase in dewpoint depression to the time of occurrance of max. temperature, but this increase does not take place so that the temperature rises quickly, but the dewpoint slowly (as was presumed), rather the dewpoint stays essentially unchanged, or is decreased slightly. At the Pápa synoptic station for example, the dewpoint depression on front-free days of the study period increased at 01 hours from 2.3° to 3.6° at 07 hours, and by the time the max. temperature occurred, the increase was to 11.9°. But at the same time, the value of the dewpoint remained essentially unchanged: The average values for 01 hours were 12.2°, for 07 hours, 12.6°, and at the moment of max. temperature, 11.9°. Under moderate conditions, the same finding is valid for the average humidity conditions of the layer between the ground and the 900 mb surface: Over Budapest, the average value of the dewpoint at 01 hours is 10.7°, but at 13 hours, 10.6°. Simultaneous with this, the average dewpoint at 01 hours on the ground was 13.0°, and this means than in the calculation of LI, one has used a value 0.7 gr/kg greater than the specific humidity prevailing at midnight and midday. Through these systematic errors, we obtained values for LI nearly 2 times smaller; of course the end result was not affected, but this is a sufficient explanation of why a surprisingly low value of LI=-3.9 was found for the limit of GI.

Even though this investigation cannot by far be considered to be thorough and final, it does throw some light on the fact that practically no error is committed if in the calculation of the afternoon starting state of an air particle, one proceeds by using the midnight ascents and in accord with the regulation, since on days

without a replacement of air mass, the fluctuations in the moisture content can be neglected.

5.4 Use of Multi-Parameter Alternative Methods to Predict Convective Thunderstorm Activity (G. Götz and G. Szalay)

From our preceeding investigations [14, 16] it was found that neither the use of a single instability index nor the use of individual quantities of state which define the hydrostatic status of the atmosphere (like e.g. the temperature and humidity conditions of the baric primary level) nor the use of quantities derived from the quantities of state (like e.g. the temperature gradient existing between the primary level or the entire humidity content of the air column) is sufficient in and of itself to define convective thunderstorm activity in all cases. The structure of the hydrostatic state of the atmosphere (as mentioned in section 5.1) is a much too involved process to permit its being called a consistent process, if we wish to base the problem of prediction on a method which, because of its one-sidedness, is often an excessive restriction of actual conditions. It may be assumed that more reliable and accurate prediction method within the limits of reality can be had, if--with better reference to actual conditions--the problem is attacked by including several parameters.

Thunderstorm research performed in the course of the last two decades led to the result that the origin of a thunderstorm is specified by factors like the stability of the air column, the moisture content available in the lower and middle layers of the atmosphere, the extent of ground warming, then by dynamic factors like forced convection, the vertical wind profile etc. If we remain in the circle of hydrostatic factors, the parameters used in the prediction (so-called predictors) must be selected in such a manner that the factors which define the static state, are among the predictors. Another basic consideration in the selection of predictors is that the values of the questionable parameter should be easily determined from aerological measurements or from the Temp-key, and there should be a close correlation with the phenomenon to be predicted (in our case, convective thunderstorm activity), and finally, a satisfactory physical foundation must exist for its application.

In the present paper, an attempt is made to work out prediction methods with several predictors belonging to the group of so-called two-category or alternative prognosis models, i.e. they have the property that by their application, the expected weather state can be put into one of two categories "a thunderstorm will form" or "no thunderstorm will form" on the basis of the present state of the air column. Preparation of this multiparameter prediction method for thunderstorm activity was attacked in a manner designed to provide information about whether it is correct to think that the actual atmospheric conditions can be better approximated if several parameters are used for the description of convective activity.

To approach the problem of conditions of convective activity in the atmosphere, we have available three parameter-groups. The first group consists of the various quantities of state and their derivants, the second group contains the instability indices, i.e. quantities which are already suitable for giving a complex characterization of the hydrostatic state of the air column and its relationship with convective activity, as described in section 5.3.2. The third group is formed of quantities in which the parameters of the first two groups are combined and the resulting quantities are then compared with thunderstorm activity. This is actually the primary task of the work which will be discussed in this section. Now if our assumption is correct, then by using identical comparison methods, we will be led to a stepwise improvement in the relationships.

As a comparison method, it seems very favorable to use the Skill-Score method of Appleman [22], which was prepared to check two-category (alternative) predictions, since the present investigation intends to deal with this type of prediction. Under such circumstances, an instability index can be considered as a prediction schematic with a parameter, in such a manner that the quantities of state and index values found within the GI, are assigned to the occurrence of convective activity, but the other values of the same quantities are assigned to the absence of thunderstorm activity.

The threshold values of SSI, LI and K corresponding to the limits of GI, were already found in section 5.3.2. Since the base quantities of state are essentially in the same relation with thunderstorm activity as the instability indices (namely, a certain value interval of the quantities of state corresponds to thunderstorm danger, and the one limit of this interval coincides with the limit of the entire value interval [16]), then in the determination of the GI limit of the quantities of state, the same principle is followed.

Now moving to the investigation of the relation existing between the above categories and thunderstorm activity, from the first category, we have selected quantities of state as the object of the present investigation: A thermal characteristic, namely the temperature of the 700 mb surface, and one hygric component of the status of the air column, namely the saturation deficite present in the same level. As thunderstorm interval, for these parameters we have the values from the paper [16]: $T_{700} \geq 3°$

or $(T-T_d)_{700} \leq 3°$

The results of a comparison with the actual, occurring weather picture are presented in the contingency tables of table 41. The values of the Skill-Score S and of the percentage probability of accuracy B are exceptionally low for all quantities of state, and for the dewpoint depression at 700 mb, S even has a negative value--a sign that none of these parameters by itself can be considered as a predictor of thunderstorm activity. It is interesting to note that the absence of convective activity (B_n) is better described by both quantities of state, than the occurrence of thunderstorms (B_j). Stated briefly, the presumption proves correct that if one pulls out a single quantity of state from the complex of elements by which the hydrostatic state of the air column is defined, then it may not be hoped to predict such a complex process as convective activity in a reliable manner. From the quantities of state, only two are picked out at random, but we have every reason to presume that even other quantities of state would not prove much better as predictors.

In the second category are the instability indices, i.e. quanti-

Table 41. Contingency Tables of Predictions Prepared on the Basis of Values of T_{700} adn $(T-T_d)_{700}$.

T_{700}	Vorhergesagt 1			
	3 Ja	4 Nein	5 Zusammen	$S = 0{,}15$
2 Beobachtet 3 Ja	66	60	126	$B = 66\%$
4 Nein	47	139	186	$B_j = 52\%$
5 Zusammen	113	199	312	$B_n = 75\%$

$(T-T_d)_{700}$	Vorhergesagt 1			
	3 Ja	4 Nein	5 Zusammen	$S = -0{,}03$
2 Beobachtet Ja 3	62	64	126	$B = 56\%$
Nein 4	72	114	186	$B_j = 49\%$
Zusammen 5	134	178	312	$B_n = 61\%$

Key: 1-predicted 2-observed 3-yes 4-no 5-together

ties by which the hydrostatic state of the air column is already defined in a complicated manner. For the present investigation, three special indices were selected from the numerous instability indices listed in the technical literature, and their relationships with thunderstorm activity have been discussed in section 5.3.2. Now let us define these relationships again by using the Skill-Score of Appleman. Naturally, no essential v new viewpoints can be expected, but in this manner, an objective comparison will be possible, and this is one of the goals of this work. As limits of GI, the values SSI = +2, LI = -3 and K = 29 are assumed.

In the contingency tables of table 42, a picture is given of the success expected when using the quantities SSI, LI and K as independent predictors for the convective activity of an air column. By using the Skill-Score values S, the important improvement is obtained over the use of individual quantities of state. Even the fact that the instability indices are complex quantities for the hydrostatic state, indicates beyond all doubt that in this manner one actually can come closer to the real factors of convective activity, and thus all indices are elevated to the series of real predictors of thunderstorm activity.

Regarding the individual behavior of the individual indices, essentially we get the same picture as in section 5.3.2: The best

predictor quantity is the SSI, and the prognostic value of this predictor is enhanced by the fact that this quantity determines equally the occurrence and absence of thunderstorm activity. But now for LI and K, the situation is different. Of these two, LI has a greater prognostic value, and this is enhanced since this index proves more suitable for predicting the occurrence of thunderstorm activity, whereas the factor K correctly describes the absence of thunderstorm activity.

Table 42. Contingency Tables for Predictions Based on Values of the Quantities SSI, LI and K.

SSI		Vorhergesagt 1			
		3 Ja	4 Nein	5 Zusammen	S = 0,64
2 Beobachtet	Ja 3	109	17	126	B = 86%
	Nein 4	28	158	186	B_j = 87%
	Zusammen 5	137	175	312	B_n = 85%

LI		1 Vorhergesagt			
		3 Ja	4 Nein	5 Zusammen	S = 0,47
2 Beobachtet	Ja 3	104	22	126	B = 79%
	Nein 4	45	141	186	B_j = 83%
	Zusammen 5	149	163	312	B_n = 76%

K		1 Vorhergesagt			
		3 Ja	4 Nein	5 Zusammen	S = 0,44
2 Beobachtet	Ja 3	86	40	126	B = 78%
	Nein 4	30	156	186	B_j = 68%
	Zusammen 5	116	196	312	B_n = 84%

Key: 1-predicted 2-observed 3-yes 4-no 5-together

Through the formation of the third category we are led to a completion of our actual, primary task: We intend to construct such multi-parameter methods of thunderstorm prediction, in which the parameters belonging to the first two categories, are simultaneously combined together. Our objective consisted in the development of such two-parameter alternative methods of thunderstorm prediction which correspond in essence to the method of Cox [24]. If our presumption is also valid, then in this manner we will come closer to the actual factors of thunderstorm activity and in this manner,

we will necessarily obtain better prediction methods.

When setting up the multi-parameter methods, the instability indices were set as the main predictors due to their complex nature. The first attempt consisted in the simplest two-parameter variation, in which two instability indices were linked, i.e. the method consisted in giving a prediction "yes" for convective activity in those cases where both of the selected instability indices lie within GI. The results obtained for the three possible combinations (SSI, LI), (SSI, K) and (LI, K) are presented in table 43. The Skill-Score values S generally underwent an increase, although the relation which can be attained with the one-parameter SSI method, was not reached in any of the cases here. It is notable that for these one-parameter prediction models, the tangible improvement in all cases consisted without exception in a more correct prediction of the absence of a thunderstorm, but the probability of an accurate prediction of the occurrence of a thunderstorm, decreased. Thus it can be maintained that by a combination of the instability indices we have two-parameter prediction methods which are highly suitable for predicting the absence of thunderstorm activity. By using them, the absence of thunderstorms can be predicted with a probability of 90%.

Another very useful, two-parameter prediction method (Tänczer [25]) was obtained by defining, in addition to the quantity SSI, the moisture index FI as follows:

$$FI = (T-T_d)_{850} + (T-T_d)_{700} + (T-T_d)_{500}.$$

Based on [25], the following values are assumed for the limits of GI: SSI = +3, and FI = 20, since according to Tänczer there can be a thunderstorm formation even at an SSI value of +3, provided that FI has a value not greater than 20, i.e. when a sufficiently large humidity supply is present in the atmosphere. According to the contingency tables of table 44, when applying this two-parameter method to the study period selected by us, an improvement in predictions over the combinations (SSI, K) and (LI, K) can be attained only when predicting the occurrence of convective activity, but the absence of thunderstorms is predicted much better by the combinations of instability indices mentioned above.

Table 43. Contingency Tables for Predictions Prepared by Means of the Diagrams (SSI, LI), (SSI, K) and (LI, K).

	(SSI, LI)	Vorhergesagt 1			
		2 Ja	Nein 3	Zusammen 4	$S = 0{,}57$
5 Beobachtet	Ja 2	92	34	126	$B = 83\%$
	Nein 3	20	166	186	$B_j = 73\%$
	Zusammen 4	112	200	312	$B_n = 89\%$

	(SSI, K)	1 Vorhergesagt			
		Ja 2	3 Nein	4 Zusammen	$S = 0{,}51$
5 Beobachtet	Ja 2	80	46	126	$B = 80\%$
	Nein 3	16	170	186	$B_j = 64\%$
	Zusammen 4	96	216	312	$B_n = 91\%$

	(LI, K)	1 Vorhergesagt			
		Ja 2	3 Nein	4 Zusammen	$S = 0{,}48$
5 Beobachtet	Ja 2	77	49	126	$B = 79\%$
	Nein 3	16	170	186	$B_j = 61\%$
	Zusammen 4	93	219	312	$B_n = 91\%$

Table 44. Contingency Table of Predictions Prepared on the Basis of Diagrams (SSI, FI). (Limit of GI: SSI=+3, FI=20).

	(SSI, FI)	Vorhergesagt 1			
		Ja 2	Nein 3	4 Zusammen	$S = 0{,}45$
5 Beobachtet	Ja 2	89	37	126	$B = 78\%$
	Nein 3	32	154	186	$B_j = 71\%$
	Zusammen 4	121	191	312	$B_n = 83\%$

Table 44a. Contingency Table of Predictions Prepared on the Basis of the Diagram (SSI, FI). (Limit of GI: SSI=+2, FI=25)

	(SSI, FI)	Vorhergesagt 1			
		2 Ja	3 Nein	4 Zusammen	$S = 0{,}61$
5 Beobachtet	Ja 2	94	32	126	$B = 84\%$
	Nein 3	17	169	186	$B_j = 75\%$
	Zusammen 4	111	201	312	$B_n = 91\%$

Key: 1-predicted 2-yes 3-no 4-together 5-observed

Now the question arises, whether it would be more correct to pose the question in reverse: It would be possible that for a value of SSI=+3 corresponding to stability where in most cases no thunderstorm occurs even for a high moisture content, but at an equilibrium situation of SSI=+2, a moisture content between FI=25 and FI=20 can lead to the formation of convective activity. The decision on this question was undertaken by using a test with the same two-parameter method, but by applying new limits SSI=+2 and FI=25. The corresponding contingency table is found in table 44a. As we see, for this new selection of limits for GI, the accuracy is improved greatly in all regards, and successful predictions are given which even exceed the success of the combination method of instability indices.

In order to discover the internal causes, the conditions occurring in the course of the study period were examined when a value of SSI=+3 occurred. We found that in this case, which lies outside of GI, there is no possibility of making a reliable decision on the appearance or absence of convective activity on the basis of the humidity conditions, since for values of FI greater than 20, the thunderstorm probability is also low, as for values which are not greater than 20. Of course, in 85% of the 13 thunderstorms occurring at this SSI value, the FI was not greater than 20, but among the 27 cases where the same value of SSI gave no thunderstorm, there was very often (in 67% of cases) an FI-value which was not greater than 20.

In the next test, 4 predictor quantities were incorporated into the prediction method. As prime predictors, we used the three instability indices, and in a manner which will allow separate consideration of results with regard to the prediction of convective activity through the individual combinations (SSI, LI) and (SSI, K). If the computed values of the index values fell in both cases into the GI, then a thunderstorm was predicted, and if the two values were outside the GI, then the absence of convective activity was predicted. The fourth predictor quantity was used in those cases where the two combinations of the instability indices gave contradictory results, i.e. the one combination led to the prediction "yes" and the other to the prediction "no". Now this fourth predictor quantity was not an

additional hydrostatic parameter; the decision in these cases was made on the basis of the configuration of the ground pressure field at 06 hours GMT. Namely, through our earlier investigations [14]--in complete agreement with the viewpoint of Cox--it was proven that the cyclonal curvature of the ground isobars (taken as an indicator for the rising air motion determined by convergence) plays an important role. Accordingly, in questionable cases, the occurrence of convective activity was predicted if a cyclonal curvature was present within the region in question in the synoptic morning chart; and an absence of convective activity was predicted in cases of an anticyclonal curvature. Whenever we faced a so-called indeterminant, isobar-free situation, then the statement "yes" was also given.

By using this four-parameter idea, a method of thunderstorm prediction was obtained which is exceptionally reliable, according to table 45. An accuracy of 88% is considered the maximum attainable limit (if we consider that the limits of the hydrostatic investigations were not exceeded this time). For predicting the absence of convective activity, this method is a help which has an exceptional accuracy (93%), but the occurrence of these storms can still be predicted with the greatest accuracy by the simplest method based on the values of SSI and LI.

Table 45. Contingency Table of Predictions Prepared on the Basis of the Diagrams (SSI, LI) and (SSI, K) and on the Curvature of the Ground Isobars

(SSI, LI), (SSI, K) 2 Isobarenkrümmung		Vorhergesagt 1 3 Ja	4 Nein	5 Zusammen	
6 Beobachtet	Ja 3	102	24	126	S = 0.71
	Nein 4	13	173	186	$B = 88\%$
	Zusammen 5	115	197	312	$B_f = 81\%$; $B_n = 93\%$

Key: 1-predicted 2-isobar curvature 3-yes 4-no 5-together 6-observed

Finally, an alternative method with 5 predictors was examined. In this case, the following quantities were selected as predictors: Two instability indices (SSI and LI), one humidity index (FI of Lebedjeva), one characteristic quantity for the equilibrium state of the lower half of the troposphere (namely, the temperature difference between the 850 and 500 mb surfaces), and again, the config-

uration of the ground pressure field. As the limit of GI, the values $FI \leq 25$ and $T_{850}-T_{500} \geq 25^{\circ}$ were assumed--based on calculations performed according to the specified principles. The predictions were prepared by means of a method which is analogous to that described above, i.e. the curvature parameter was applied in those cases where the combinations (SSI, LI) and (FI, ΔT) led to opposing predictions on the occurrence of convective activity. According to table 46, a value of the Skill-Score was attained which is 0.02 less than in the case of the preceeding 4-parameter method, but this scheme proves to be highly suitable for predicting the occurrence of thunderstorm activity. The decrease in the Skill-Score value again comes from the wrong prediction of the absence of thunderstorms, but with this 5-parameter method (which is still fast to use) the periods without storms are still predicted with an accuracy of 87%.

Table 46. Contingency Table for Predictions Prepared on the Basis of the Diagrams (SSI, LI) and (FI, $T_{850}-T_{500}$) and on the Curvature of the Ground Isobars.

2 (SSI, LI) (FI, $T_{850}-T_{500}$) Isobarenkrümmung		Vorhergesagt 1: 3 Ja	4 Nein	5 Zusammen	
6 Beobachtet	Ja 3	111	15	126	$S = 0{,}69$
	Nein 4	24	162	186	$B = 88\%$
	Zusammen 5	135	177	312	$B_f = 88\%$
					$B_n = 87\%$

Key: 1-predicted 2-isobar curvature 3-yes 4-no 5-together 6-observed

Briefly, we find that the investigations give a complete justification of the assumption that the description of convective activity of an air column even within the circle of hydrostatic parameters, has to be examined by the use of multi-parameter methods, especially since they are better suited for approximating the actual, complicated conditions. The alternative methods presented in this section are only a part of the possibilities available in this direction. It was not our intention to examine all the possible combinations by variation of predictors. We viewed our task only as a presentation of some viewpoints which are basic to the examination of the question of convective thunderstorms. The presented prediction methods serve primarily to support these viewpoints.

5.5 Absence of Convective Activity in an Air Column in Instable Equilibrium States (G. Götz and G. Szalay)

The main emphasis of previous sections was directed at factors by which the structure of convective activity is promoted, i.e. the conditions of the hydrostatic state were examined from the area of the occurrence of thunderstorms. This relatively one-sided view arises from the essence of this question: It comes from the considerations of the different Weather Services, and primarily from the View of the Storm Warning Service at Lake Balaton, where we are dealing largely with so-called precautionary predictions, i.e. the correct prediction of the elements of "bad weather" have a much greater importancethan the prediction of "fine weather". From a theoretical viewpoint, there is no such distinction, and, as already mentioned, the predictions of a disturbed weather picture and of fine weather are theoretically equally valuable.

If we overlook subjects treated in the preceeding sections, then two previously untouched questions come up: What are the reasons for the occurrence of convective activity in a stable air column and for the absence of it in an instable air column? The answer to the first question belongs entirely to the circle of dynamic research, namely, in such cases, the different types of forced convection (frontal or orographic lift, lift caused by convergence of the wind field etc.) are active. The second question is two-fold: There are some dynamic and some hydrostatic reasons which can lead to the absence of thunderstorm activity, even under thermally instable conditions. A thorough investigation of these factors--although they are usually pushed into the background due to the mentioned circumstances--is theoretically no less important and serves in the final analysis, in a practical manner in thunderstorm prediction. The dynamic reasons are the forced, vertical motions (which can be frontal, orographic or convergence-induced), and the hydrostatic causes in the vertical distribution of quantities of state.

The absence of thunderstorm activity in instable layering is attributed to two factors in this report; these are descriptive of the air state. The one factor consists in the lack of available moisture, the other factor is the insufficiency of near-ground warming

due to insulation. The restrictive or preventive effect of these two factors on thunderstorm activity is well-known; the data below will provide a certain insight into the extent of these influences.

The equilibrium state of the air column is considered instable when the values of the individual instability indices indicate an instability. As limit values for the lability, values of SSI=+2, LI=-3 and K=29 were selected. To characterize the humidity conditions in the lower layers of the troposphere, the elevation of the condensation level was computed for an air particle which begins its movement at a level of 850 mb. To identify the insulation warming, the potential temperature was used which corresponds to this lifting condensation level. According to our investigations [16], 75% of the convective thunderstorm activities occur in those cases where this condensation level is below the 760 mb surface. Based on this, it was presumed that the lack of moisture in such cases can lead to a restriction of convective activity, in which the condensation level lies at a level of 760 mb or more, i.e. when the dewpoint depression at the 850 mb surface is 7 degrees or more. However, the size of near-ground heating was then considered insufficient for the formation of convective thunderstorms when the max. temperature of the studied region was not higher than the potential temperature of the condensation level.

According to our investigations, in an instability based on SSI, in 67% of all cases, the absence of thunderstorm activity did not have the necessary values of ground temperature. A too-high condensation level (lack of moisture in the lower air layers) occurred in 15% of cases, but in all these cases, an insufficient ground warming was present, and the low humidity never occurred alone. In the evaluation of this fact, it must not be forgotten that the values of the SSI already contain certain elements of the humidity relations of the air column (or more accurately, they contain the saturation deficit at the 850 mb surface), and thus in a too-dry air column, they give inherently no instability. Consequently, the fact that the thunderstorm activity expected because of the SSI did not occur in 15% of the cases (at least in part) due to the lack of moisture, indicates that this index gives a correct evaluation of the moisture content of

an air column, or that the moisture conditions in the vicinity of the 850 mb surface play an important role in the generation of convective activity.

In an instability determined on the basis of the factor K, the absence of thunderstorm was accompanied in 86% of cases by an insufficient ground warming, and in 17% of cases, a lack of moisture was present, again simultaneous with the insufficient warming. The moisture conditions of the air column are thus correctly reflected by the factor K (this is assured by the dewpoint at the 850 mb surface and by an approximate value of the humidity in the layer below the 700 mb surface).

For LI the situation is quite different. Here the insufficient ground warming was the cause of the absence of thunderstorm activity in only 52%, and this indicates that this index contains the predicted value of the temperature maximum on the ground. Conversely, a too-high position of the condensation level occurred in 33% of all cases, and in 9% of cases, this was the sole active factor. From this it can be concluded that an exclusive consideration of only the humidity conditions of the lowest air layers (as is done for LI) is not always useful; layers are very important to the process of thunderstorm formation.

Concerning the role of the lack of moisture one can get a more accurate picture if this question is considered not in the light of the instability indices. For the study period, in all cases where a convective activity did not occur, the values of FI were determined. According to calculations, in half of these storm-free cases, the value of FI was greater than 20, and in 86% of cases, it was greater than 10. Conversely, only 33% of thunderstorms belong to FI values greater than 20. Due to this fact, some light is thrown onto the circumstance that the lack of moisture might play a role in the absence of thunderstorm activity, but this case is not too frequent--at least not around Hungary--so in a consideration of all cases in the study period, 56% had an FI value less than 20, and 90% had values less than 30. Thus, the number of cases is relatively small in which a significant lack of humidity is actually present within the advected air column.

5.6 Some Considerations of Kinematic Character in Connection with the Outbreak of Convective Thunderstorms (P. Ambrózy)

It is apparent that the outbreak and course of a convective thunderstorm are affected by the prevailing thermal conditions and by numerous other factors. One of these factors, or a group of them, can have such a severe effect that in spite of the otherwise favorable structure of thermal characteristics, the thunderstorm formation is suppressed. One such factor is the vertical wind shear, i.e. the sharp change in the wind vector with elevation.

For a more detailed investigation of this problem, it is helpful to study the relation between the high-altitude wind and the convective thunderstorm activity.

As is known, convective thunderstorms occur with greater frequency in weather situations having an indistinct or weakly gradient thermobar field than in areas characterized by a pronounced flow of high speed. To illustrate this fact, table 47 shows the average winds at various altitudes from the Summer months (May to August) for the years 1962 and 1963, based on Budapest ascents at 12 hours GMT, separately for days with and without thunderstorm. As we see, at all altitudes there is a pronounced difference in which the high-altitude wind seems generally weaker in the time before a thunderstorm.

Table 47. Average Wind Velocity in Budapest on Days with and without Thunderstorms (May-August 1962-1963, 12 hours GMT)

1 Höhe	2 Ohne Gewitter	3 Mit Gewitter	4 Differenz	5 Differenz ausgedrückt in Prozenten der Werte der grewitterfreien Tage
500	5,8	5,4	0,4	7
1000	6,0	4,7	1,3	22
1500	6,5	5,2	1,3	20
2000	7,7	5,7	2,0	26
3000	10,1	7,2	2,9	29
4000	11,7	7,5	4,2	36
5000	14,3	8,7	5,6	39
7000	17,8	12,0	5,8	33
9000	20,9	16,3	4,6	22

Key: 1-elevation 2-without thunderstorm 3-with thunderstorm 4-difference 5-difference expressed in percent of values without thunderstorm

A similar picture is also given by the distribution of the vertical wind shear. The wind shear vectors were determined between all altitude steps represented in table 47 (beginning at 1000 m elev-

ation) and for all wind measurements taken in the Transdanube basin and in Budapest in the course of the study period. We formed averages for days with and without thunderstorms. In this manner, we obtained average values on days without thunderstorms of 9.6 m/s/1000 m, and for days with thunderstorms, 6.7 m/s/1000 m. It is worthwhile to note that on days in which one of the lability indices (i.e. the index of Showalter, the factor K, or the humidity index) indicated a pronounced probability of thunderstorms, but when no thunderstorm actually occurred, the average wind shear was 7.6 m/s/1000 m, or somewhat higher than on days with thunderstorms. We will return to this question again later. Since the above values were determined on the basis of data on the high-altitude wind measurements in the Transdanube basin (and in Budapest), our conclusions will also apply for the Balaton region which assumes the central position in the high-altitude wind observations.

Now we can move on to an explanation of the question of how the influence of wind shear is expressed in cloud and rainfall formation.

At first glance, the fact seems to contain a contradiction, that the vertical wind shear can both promote and retard cloud formation. We will first examine the former question.

Besides thermal turbulence--and besides the more orderly motion of the thermal upwind--the dynamic turbulence can have a part in the formation of convective cloud formation. The "heat equivalent" of this factor can be expressed by the following equation [27]:

$$\gamma_d = \frac{T}{g}\left(\frac{dV^2}{dz}\right)$$

whereby T is the average temperature of an air layer of elementary thickness dz, g is the acceleration of gravity and dV/dZ is the vertical wind shear. This dynamic turbulence occurs on the side of the convection and opens up the possibility that in the case of a sufficiently large wind shear, a rising air motion and cloud formation can occur even for a vertical temperature gradient which is significantly smaller than the adiabatic gradient. In this manner, SC, ST or even Cu clouds can be generated, usually with a lower cloud base, but with little vertical thickness. This case occurs

primarily after powerful cold outbreaks on the ground. (In order that the component γ_1 should reach the value 0.3-0.4, in the air layer of 1000 m thickness, a vertical increase in the horizontal wind of 10-15 m/s must be present).

A wholly different effect is exerted by the vertical wind shear on the clouds generated by thermal convection with vertical organization in those cases where the wind shear occurs above the cloud base. If the horizontal air motion existing within the cloud (which is also present outside the cloud) increases with elevation, then a distortion of the cloud occurs in the direction of the wind shear vector. If this factor occurs in a very pronounced manner, then the further development of the Cu cloud into a Cb can be prevented. But the wind shear still has a great influence on the structure and total life of the cloud, when the amount of wind shear is not sufficiently large to prevent the formation of a thundercloud.

Since in the cloud a significant wind is present, i.e. a vertical pulse flow is met, then the value of wind shear inside the cloud is less than in the environ. The distortion of the cloud is thus not as great as could be expected on the basis of high-altitude wind measurements.

The cloud-distroying effect of wind shear depends on its ability to predominate over factors promoting cloud formation: These are the speed and mass of the updraft. If the latter are small, then the cloud is broken up on entry to the zone of wind shear. In the case of an updraft extending over a larger area and having a great speed, the cloud can penetrate far into the distortion zone without being cut up from the energy source below, causing cloud dissolution. According to Byers and Braham [26], a thunderstorm can even form when an average wind shear of 30-35 km/h is present in the layer between the 850 and 300 mb surfaces. We also encountered similar values in convective thunderstorms.

Under consideration of these facts, the question of cloud-destroying effects of vertical wind shear was examined in the Trans-

danube Basin. The investigations could not be held within a smaller space, like the Balaton region, due to the lack of data caused by the irregularity and small height range of pilot balloon measurements.

Our attention was turned primarily to the question of whether the absence of thunderstorm formation in otherwise favorable lability conditions was a result of vertical wind shear. In the years 1961-1964, there were 28 days when the prerequisites for outbreak of a thunderstorm were present due to the lability indices (i.e. SSI had a value smaller than -1 or the factor K had a value greater than 30), but nevertheless, no thunderstorm activity occurred. Of the 28 cases, only in 2 of them can it be assumed that the absence of thunderstorm activity was caused by wind shear, and in about 7 cases the wind shear probably was only a contributing cause with other factors which prevented thunderstorm activity (in 25% of all cases). Conversely, on 100 days with convective thunderstorms, in 32 cases (32%) there was a wind shear of the same order as on the 7 storm-free days. Due to a lack of information on the speed and size of the ascending air flow, it cannot be determined how much the intensity and spatial extent of the thunderstorm could have been reduced by wind shear. At any rate, the conclusion can be drawn that the presence of a wind shear may not be considered as a phenomenon which will prevent the generation of a thunderstorm, with the exception of the case where several other factors act decisively against thundercloud formation.

We also examined the question of whether in cases with strong wind shear, the average amount of rainfall is smaller or not. This investigation could only be performed for the Budapest region, since high-altitude wind measurements in the Transdanube basin reach only a low height, and consequently too little data was available. In consideration of the fact that high-altitude wind measurements are valid not for a given point, but for a larger region, from the environ of the Budapest-Lorinc observatory, where the measurements were taken, we have the averages of precipitation at the Stations of the Budapest Meteorologic Institute: Gödöllo, Nagykáta and Orkény; we then compared these averages with the wind shear. Contrary to expectations, we found that on days with wind shear, the averages of precipitation were just as high as on days without wind shear. The

convection-destroying effect of wind shear thus was not felt. But it must be noted that the number of cases handled in this study (68 cases without wind whear, 32 cases with wind shere) was not large and causes the average values of convective precipitation--which are known to vary widely--to be somewhat unreliable in spite of areal averaging.

Since the outbreak of a thunderstorm and its intensity are hardly affected by wind shear, other factors must be taken into account if the reasons responsible for the failure of thunderstorm outbreak in otherwise favorable lability conditions, are to be found. In the majority of the questionable 28 cases, this cause (or these causes) were quickly found. In 10 cases, it was the low moisture content in the middle troposphere, in 9 cases the weak development of convection-causing condensation (near-ground temperature did not reach the necessary temperature), and in two cases, both causes were responsible for the absence of thunderstorm activity.

It is equally useful to examine the thermobar fields of these situations. We found that in more than half of the cases, a high-altitude anticyclone and a warm-air region were present over the Carpathian basin. In such synoptic situations, there is often a falling air flow and this factor also acts against convection.

The advection relations on thunderstorm days also presents an interesting picture (including days without thunderstorm, but with a lability; SSI less than +1, K greater than 30). These values are found in table 48, computed on the basis of the vertical changes in wind vector between the altitudes 1000/3000 or 1000/5000 m, which were formed by averaging from the data of high-altitude wind measurement stations in the Transdanube basin for the summer months of the years 1960-1964.

It is surprising that in cases where the thunderstorm activity was absent, the frequency of cold advection into the lower troposphere was significantly higher. The cases of cold advection exhibit no pronounced similarity in structure of the thermobar field. In the layer 3000/5000 m, no difference can be found in the frequency of warm advection and cold advection between cases with and without

Table 48. Percentage Frequency of temperature Advection

Advektion 1	3 Warm		6 Kalt		7 Unbestimmt	
	2 Mit	Ohne 4	2 Mit	Ohne 4	2 Mit	Ohne 4
	Gewitter 5		Gewitter 5		Gewitter 5	
1000/3000 m	51	32	16	45	33	17
3000/5000 m	37	36	41	42	22	21
1000/5000 m	51	29	26	54	23	17

Key: 1-advection 2-with 3-warm 4-without 5-thunderstorm 6-cold 7-undefined

thunderstorm. The advection computed for the layer 1000/5000 m is under the powerful influence of the conditions prevailing in the lower layers.

In the interest of completeness, for Siófok, the percentage frequency distribution of wind directions was determined at the upper limit of the friction layer (1000 m) before the outbreak of the thunderstorm (at 12 hours GMT). The results are in table 49. In the interest of interpreting the data, the available, average wind direction frequencies for Siófok, for 1000 m altitude, are also reported (from the years 1954-1958).

Table 49. General, Average Wind Direction Frequency and Wind Direction Frequency on Thunderstorm Days in Siófok

	N	E	S	W
Thunderstrom days	34	17	28	21%
All days	32	11	33	23%

As we see, with the exception of the East, there is no difference in frequencies on days with thunderstorms and the other days. For an East wind, there is an increase in thunderstorm probability. This is apparently a result of the well-known meteorological fact that the East weather situations in the Summer months are connected with a marked thunderstorm tendency.

6. Instability Lines and Potentials for Their Prediction in the Area of Lake Balaton (I. Bodolai and E. Bodolai)

The most powerful storms on Lake Balaton usually occur in connection with instability lines and thunderstorm cold fronts. The knowledge we have about instability lines came from the mesosynoptic analysis. At this point, we will not discuss the structure of the mesosynoptic analysis nor its methodology. Several papers have been

published on this subject in recent years [7, 28, 29, 30, 31]. We offer here only a brief summary of the concept of the instability line and some important properties of these lines.

The concept of an instability line (from the original English name Squall, Squall-line, and from the German designation Böe, Böen-Line [squall, squall-line]) has long been in use in synoptic meteorology. But this concept has undergone an important change: In the period before the detection of the fronts, the actual fronts are called squall lines etc. and after detection of the fronts, the instability lines are called cold fronts. Today, by instability lines we mean a region where the convective activity occurs in an orderly manner along a line; this line is usually a strip of 50 to 80 km width and the front side of the strip forms the actual instability line. This line is accompanied on the ground by line-like thunderstorms having strong wind gusts, hail and sometimes tornadoes. Instability lines are usually found on the South side of young cyclones within the warm sector, East of the cold front. The weather events accompanying the lines resemble in many respects, cold fronts. But between these two synoptic objects there is the basic difference that after passage of a cold front, due to the onset of air-mass replacement, the meteorological elements retain their changed values for a longer time, whereas after passage of an instability line, the meteorologic elements have those values existing before passage of the line.

In mesosynoptic analysis the instability lines appear as objects having a peculiar structure and separate development history. Characteristic mesosynoptic objects connected with the instability line are:

1. The thunderstorm-high, i.e. usually an elongated, high-pressure feature located beneath the cloud chain of the Cumulonimbus.

2. The zone of positive pressure change or of the pressure jump; a zone of large pressure gradient lying on the front side of the thunderstorm-high, and it is actually the leading edge of this high, called the squall line, or using international terminology, the instability line. Along this line, the jump-like changes in the

other meteorologic elements take place.

3. The zone of negative pressure change is a zone with large pressure gradients located behind the thunderstorm-high. One difference with the above zone of positive pressure change can be that the transition from the thunderstorm-high into the region of lower pressure is not so sharp and jump-like, than in the zone of positive pressure change. Consequently, this zone assumed a greater width on the mesosynoptic map.

4. The loop-depression is a small cyclone located behind the zone of negative pressure change. This mesosynoptic object is generated similar to a whirlwind forming behind a solid object moving on the ground (imagine a car moving along a highway, with the role of the moving, solid body taken over by the air mass of the thunderstorm high). This object usually only appears in a later stage of development of the instability line, and sometimes it does not form at all.

During the dissolution stage of the instability line, a stepwise breakdown of the thunderstorm-high occurs; the amount of the pressure jumps becomes ever smaller, and the loop depression is filled in stages.

In researching the structural peculiarities of the instability lines in the course of the last decade, important progress has been made. But the generation of these lines and circumstances surrounding their origin are little-known. This is largely due to the unfortunate fact that the temporal and spatial density of aerological ascents is insufficient to allow a more accurate discovery of mechanisms belonging to such, smaller dimensions. According to our present conception, the most frequent cause of generation is found in the distribution of temperature advection after the high. This distribution is favorable for the formation of an instability line if an intensive heat advection is present in the lower layers, while in the middle and upper troposphere, a cold advection is underway, thereby in a quite thick troposphere layer, a concentration of potential instability can occur. This mechanism occurs often in the warm sector of young cyclones in which a quite sharp temperature

contrast forms in the frontal zone of the middle troposphere. But this conditions is not the sole genetic cause (although it is met rather often) for the formation of an instability line, since other circumstances, like e.g. a strong convergence line forming in the near-ground layer, can likewise exert a favorable influence on the generation of an instability line.

The above discussion contains only a summary of the most general indications of an instability line. But from this brief overview we can extract the fact that it is possible to discover the instability lines by means of a mesosynoptic analysis, to track them in practice within the Storm Warning Service. Since the strongest instability lines occurring over the last 10 years had their mesosynoptic properties analyzed in detain in the papers [7, 8], we shall present below the outlines of some important considerations regarding their prediction.

An analytic method for predicting instability lines is not available. Since an inertial extrapolation of meteorologic fields is insufficient for a determination of the various phases of formation, the prediction of these fields can be performed only by synoptic means. For this reason, special attention must be given those synoptic situations which represent the most favorable circumstances for the formation of the instability lines.

Two examples for the two most-typical synoptic situations are presented below, in which the instability lines formed within the Carpathian basin. All instability lines of the last 10 years which were already subjected to a mesosynoptic investigation, occurred in weather situations belonging to one or the other of these two main types [8].

Figure 65 has maps illustrating cases where the instability line was formed within a regional cyclone. A characteristic example for this is the case of 3 June 1958. The instability line visible in fig. 65a moved along the near-ground convergence line of a regional cyclong lying over the Alps. The near-ground cold front extended along the Czechoslovak-Austran border, and became a warm front on the South edge of the Alps, which connects with the pressure minimum

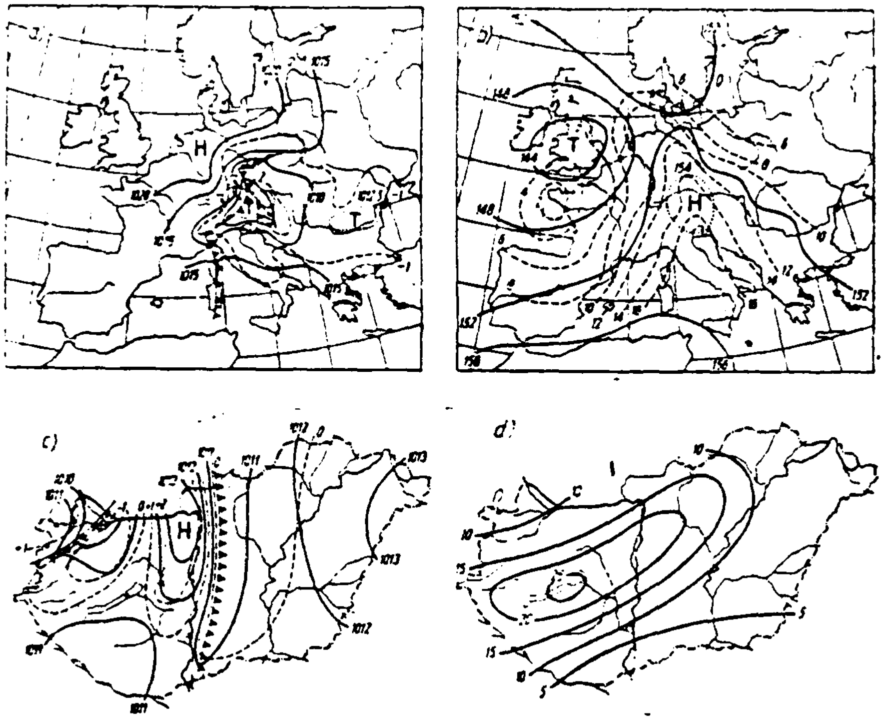


Fig. 65a) Weather Situation on 3 June 1958, 15 hours GMT; b) Absolute Topography at 850 mb on 3 June 1958, 00 hours GMT. The solid lines are isohypses, the dashed lines are isotherms; c) Mesosynoptic situation on 3 June 1958, 21 hours GMT. Solid lines: Isobars; Dashed lines: ISallobars; d) Distribution of Max. Wind Gusts (m/s) on 3 June 1958.

in the Po valley, and then continues with the property of a cold front to the South over the bay of Genoa. In the case of the instability lines occurring in this weather situation, the outermost Northern extent of the cold front does not go beyond the 52nd parallel and it lies in a N-S direction between the 13th and 15th medidians.

The topography of the 850 mb surface at a time 15 hours earlier than the time of the ground map, is seen in fig. 65b. The thermal field of the isobar surface is quite sharp. The frontal zone extending over Western Europe has a N-S orientation. The warm backside is moving from South Italy over the northern Adriatic and the Alps in the direction of the North sea. The midpoint of cold air is over West France. Notice that the topography of the 700 and 500 mb surfaces give a similar picture. It is notable that 12 hours before the formation of the instability lines, the axis of the warm backside at the 850 mb topography is always found over the Carpathian basin.

In this example, the instability line discernable at 15 hours GMT reached the border at 16 hours GMT, at 18 hours GMT it moved over Siófok and caused a wind gust there of 25 m/s.

The structural peculiarities mentioned above for the instability line are clearly seen in fig. 65c. The spatial location of the line, the following zone with large gradient values and the thunderstorm-high can be precisely analyzed. In this case, we find that the loop cyclones flow together behind the thunderstorm-high, with the prefrontal pressure drop of the following cold front. As we see in fig. 65d, after passage of the line, there were wind gusts in a large part of the country with exceeded 10 m/s. It should be mentioned that the instability lines belonging to this group can have, in many cases a significantly more powerful development and give wind gusts over 30 m/s in the Lake Balaton region.

The second synoptic situation favorable for the formation of an instability line, was shown in fig. 66. The case of 10 June 1960 is presented, where the cold front connected with the instability line goes out from the edge of a cyclone over South Scandinavia and extends down from there is a southerly direction. For the synoptic situation in question, the cyclones are found in the area of the North Sea and South Scandinavia, the northern limit of the cold front is located on the 55-56th latitude and between the 10th and 15th meridians in a southerly direction. At the South slopes of the Alps, the cyclones have, in all cases, a continuation in the form of a warm front (similar to the above weather situation) and open into a depression in the Po valley. In this weather situation, the longitudinal extension of the cold front is significantly greater than in the case of the regional cyclones. The majority of the instability lines is generated in this case, in the area of the Istrian peninsula and Slovenia.

In the topography of the 850 mb surface from the time 12 hours earlier than the time of the above weather situation (fig. 66b), we see a similar structure of the thermal field as in the previous weather situation, only the position of the frontal zone in this case has a greater deviation from the N-S direction. The axis of the warm

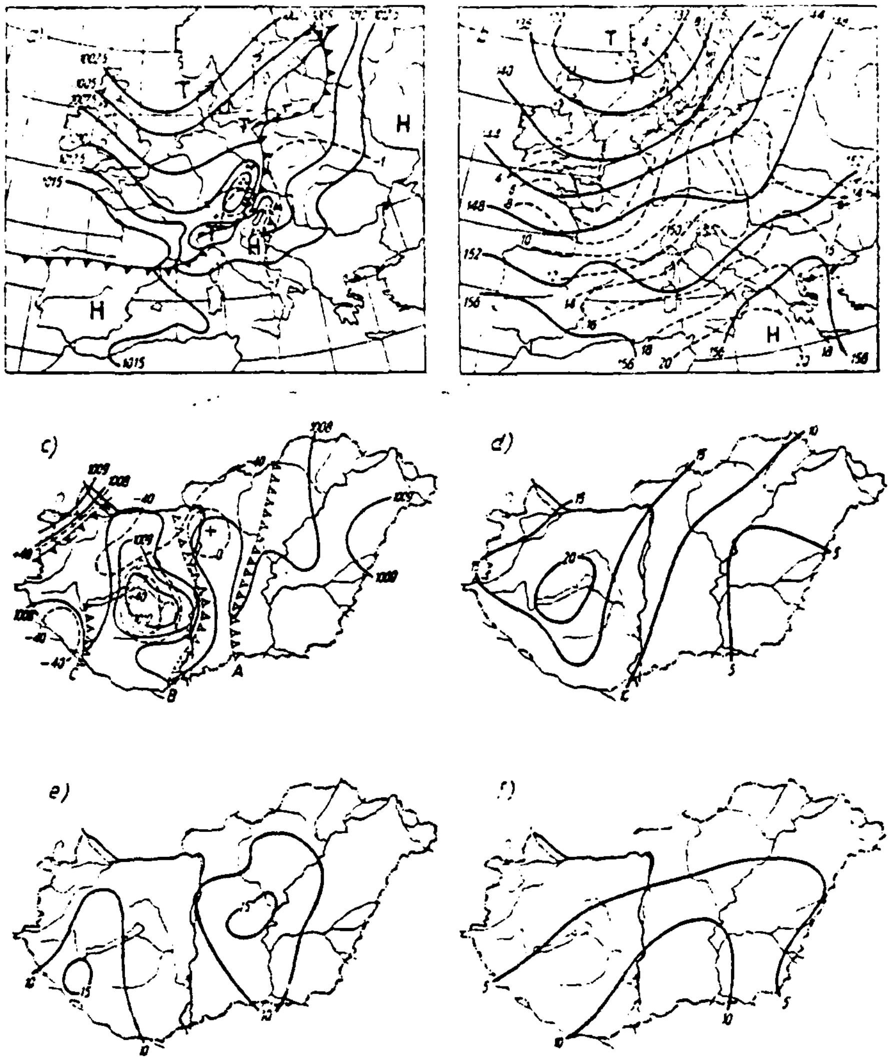


Fig. 66a) Weather Situation on 10 June 1960, 12 hours GMT; b) Absolute Topography of 850 mb on 10 June 1960, 00 hours GMT (Designations as in fig. 65/b); c) Mesosynoptic Situation on 10 June 1960, 18 hours GMT (Designations as in fig. 65/c); d) Distribution of max. Wind Gusts (m/s) occurring after Passage of the Instability Line A; e) Distribution of max. Wind Gusts (m/s) occurring after Passage of Instability line B; f) Distribution of max. Wind Gusts (m/s) occurring after passage of Instability line C.

backside is found over the Carpathian basin, approaching along the course of the Tisza river. As already mentioned above, this circumstance is characteristic for all cases in which a time difference of about 12 hours exists between the times of preparation of the topography and the situation forming instability lines. At the

place where the formation of the instability line is proceeding by itself, in the middle troposphere in both synoptic situations, a cold advection is underway. In the present case, there was a formation of three sequential instability lines, which are visible on the mesosynoptic map of fig. 66c. The lines designated B and C are generated in the Southwest part of the Transdanube basin. The latter situation is also an example that the instability lines in the Carpathian basin occur in a changing mesosynoptic structure, and powerful storms can follow in sequence at relatively short time intervals, even within hours, as is illustrated in figs. 66d and 66e on the basis of max. wind gusts generated by three instability lines.

The synoptic situations illustrated by the above examples are difficult to predict due to their quick formation, and just for this reason it is very important that in such macrosynoptic situations, the analysis of the ground weather charts must be performed with the greatest thoroughness and care. In questionable weather situations it is always important to dedicate great care to the beginning of convective activity in the Southeast zones of the Alps. If a synoptic situation is found which promotes the formation of an instability line, and if the onset of convective activity is recognized in time by means of a mesosynoptic analysis, then on the basis of knowledge of the average behavior of the instability lines, a prediction of this dangerous synoptic object can be performed in the frame of the Storm Warning Service.

From the viewpoint of predicting the instability lines, it is advantageous if we have available information concerning the average values of the meteorologic elements accompanying the passage of such a line. In table 50, the following data is compiled for the previously observed lines: The average transit time over the country; the time in which the lines move from the border to Lake Balaton; the temperature and wind jumps occurring in the wake of the thunderstorm, and all these as a function of the velocity categories presented in the vertical columns of table 50. For a comparison and to throw some light on the problem of such predictions, the same data was cited for the thunderstorm cold fronts; the latter was taken from

our earlier work presented in this manuscript. By a thunderstorm cold front, we mean that thunderstorms are distributed linearly along it.

Table 50. Some Average Characteristics of the Streaming of Instability Lines (Il) and Thunderstorm Cold Fronts (Kf_g) in the Case of Strong, Moderate and Weakly Developed Synoptic Objects.

	1 Stärke des Sturmes m/sec	2 Sturmart	3 Zeitdauer des Durchzuges (St.)	4 Die für das Erreichen des Balatons nötige Zeitdauer (St.)	5 Temperatursprung beim Gewitter (°C)	6 Mittlere Maximalgeschwindigkeit (m/sec)
7 Stark	$V_{max} \geqq 30$	Il	6,8	0,9	—13,0	35
	$V_{max} \geqq 25$	Kf_g	9,7	2,4	— 5,4	29
8 Mässig	$V_{max} \geqq 25$	Il	7,1	1,2	— 9,6	27
	$V_{max} \geqq 20$	Kf_g	9,4	2,7	— 5,9	22
9 Schwach	$V_{max} \geqq 20$	Il	7,0	0,8	— 5,8	21
	$V_{max} \geqq 10$	Kf_g	9,3	3,0	— 4,0	17

Key: 1-storm intensity 2-storm type 3-duration of passage (hours) 4-time in hours, needed to reach the Balaton area 5-temperature jump for the thunderstorm 6-average, maximum velocity 7-strong 8-moderate 9-weak

From table 50 we see that in the case of the instability lines, there are important differences between the three wind speed categories in the values of the temperature jumps, but for the thunderstorm cold fronts, there are no significant differences between the various wind speed categories. From a comparison between the characteristics by which the two synoptic objects are identified, we can see that the thunderstorm cold fronts are in every respect a weaker formation than the instability lines. The upper limits of the speed categories for thunderstorm cold fronts are 5 m/s lower, and the lower limits are 10 m/s lower than for the instability lines. From the transit times we see that the movement of the thunderstorm cold fronts is much slower since they reach the Lake Balaton region on average in 2.7 h, compared to 1-1.5 h for the instability lines. From this we can suppose how difficult it is to predict instability lines, and that a successful prediction of them is impossible without mesosynoptic information.

For all the weather situations illustrated above, the exception-

ally rapid development is characteristic. Today, we are not yet able to predict such rapid developments by analytic methods. In spite of the difficulties, in the United States, certain attempts were made in this direction. The prediction methods known to us have an empirical-synoptic character. In the book by George [32], the methods are reported which were used in various regions of the US to predict instability lines. These methods are based on the determination of the synoptic situations which are favorable for the formation of an instability line. One of these empirical prediction methods can be used in the case of the classical prefrontal instability lines and will be discussed briefly here and applied to some instability lines. The method is based on the use of 12-hour isallohypsis charts, the 850 mb surface and the moisture field at the 850 mb surface. Even though this method in used in the USA only in the period from November to April, we tried to apply it to the summer weather situations. This can be justified since the method has the appearance of being independent of the geographic situation and since values of the temperature contrasts occurring in summer here resemble those occurring in the winter in the USA.

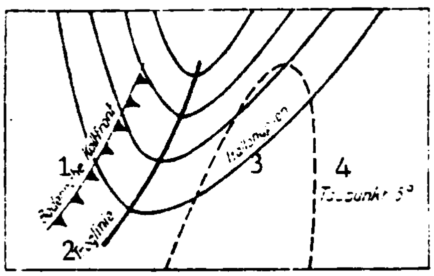


Fig. 67. Opposing Situation of the Isallohypses, the moist air tongue, the Trough Line and **near-** ground Front in the Case of a Small Convective Activity (from George [32]).

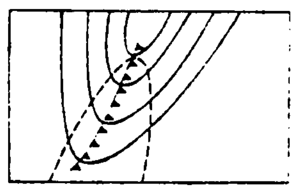


Fig. 68. Outline of the Convective Activity Occurring behind a Front (from George [32]).

Key: 1-near-ground cold front 2-trough line 3-isallohypses 4-dewpoint

The method of George is broken down into three phases: 1. The determination of the weather situation which promotes the formation of an instability line; 2. Statement of the time of formation; 3. Determination of the location of this formation. According to George, the 12-hour isallobare chart and the humidity field at 850 mb can provide four different variations with respect to the formation

of an instability line. These 4 variations are obtained by drawing in the trough line of the isallohypses for the region in question at the isallohypsis chart of 850 mb, in addition, the near-ground fronts and the isoline of the 5° dewpoint are entered. The following 4 cases can result on the processed chart:

1. In cases illustrated by fig. 67, only one small frontal or prefrontal convective activity can be expected, since both the cold front and the axis of the trough line are located behind the tongue of moist air.

2. If the near-ground cold front lies in the moist tongue (fig. 68), then a prefrontal convective activity is not expected; it will not occur until after the front.

3. Situations sometimes favorable for the formation of an instability line are shown in figs. 69 and 70. In fig. 69 we have a trough line which differs from the front only in its Southern parts. In such cases, the potential for formation of an instability line does exist, but it will be weak and hard to find. In the case of fig. 70, the formation of an instability line is expected only in the section of the trough line having a negative sign.

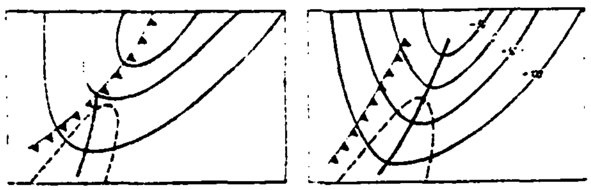


Fig. 69-70. Opposing Situations, Sometimes Favorable for the Formation of an Instability Line (from George [32]).

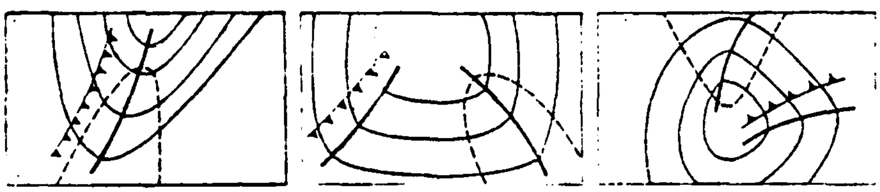


Fig. 71-73. Opposing Situations Which are Highly Favorable for the Formation of an Instability Line (from George [32])

4. In figs. 71, 72, and 73 we see the situations which are highly favorable for the formation of an instability line. The axis of the trough line has various configurations compared to the near-ground

front which is located within the moist tongue.

If we use the above variations to find that the situation is favorable for the formation of an instability line, then the question must be decided of when and where this synoptic object will be in the course of the next 12 hours. For the second question, an answer can be found in the above outlines: Namely, the instability lines appear within the moist air tongue along the trough line of the isohypses. Concerning the propagation rate of the trough line, according to American investigations, the following regularity exists: If the trough line is located 240 km to the East of the near-ground front, or if this distance is even less, then the propagation of the trough line occurs at the same speed as the front; but if the trough line is more than 240 km East of the front, then the trough line remains stationary until the time when the separation declines to 240 km.

It is difficult to give the timing for the formation of an instability line since an important time difference may exist between the first appearance of the weather situation favorable for the formation of the line, and the first actually analytic appearance of the line. To estimate this time difference, the following empirical rule can be used:

1. If the average separation between the trough line of the isallohypses 850 mb and the trough line of the 500 mb surface is more than 1500 km, then the formation of the instability line occurs with a delay of 21 hours over the first finding of a favorable high-altitude weather situation.

2. If the mentioned distance has a value between 650 and 1500 km, then the appearance of the instability line occurs with a delay of 15 hours over the timing of the particular isallohypses chart.

3. If the distance is less than 650 km, then the appearance of the instability line can be expected in about 9 hours.

Six hours can be deducted from the above delay times if the moist tongue is located in the Eastern half of a closed isallohypsis focus.

The described prediction method is valid for the classical prefrontal instability lines, as mentioned above. But now the formation of an instability line is in many cases linked to a specific geographic region. This is the case for instability lines frequently encountered here, since the majority of them form in a specific geographic zone, on the Istrian peninsula and in Slovenia, and within the warm sector of an edge cyclone located over the Po valley and in regions in front of the cold front of this cyclone. The instability line is first passed by the other cold front over the Carpathian basin--this latter front belongs to a cyclone lying over the North Sea or Denmark. Thus, the formation of these special instability lines occurs not in the classical manner, but under the joint influence of the frontal effect and the favorable conditions prevailing in the North Adriatic region. Precisely for this reason, the method described above can only be used here for test purposes on the instability lines occurring in our country, and such a test can only have a diagnostic significance. Under the climatic and geographic conditions prevailing here, only by studying many cases will it be possible to prepare reliable, prognostic rules.

The attempt at prediction was performed on 4 instability lines based on 24-hour isallobars: These were the instability lines of August 12 and 19, 1960, July 13 and August 8, 1963.

These cases all belong to the second-type weather situation. If we look at the charts of fig. 74, then we see that these situations all have a similarity with the outline of George's fourth variation illustrated in figs. 72 and 73. The trough line crossing the Carpathian basin, is found within the moist tongue, and even though the South cold front likewise coincides with the trough line, with regard to the dryness of the air, no prefrontal convectivity activity can be expected in front of this frontal section.

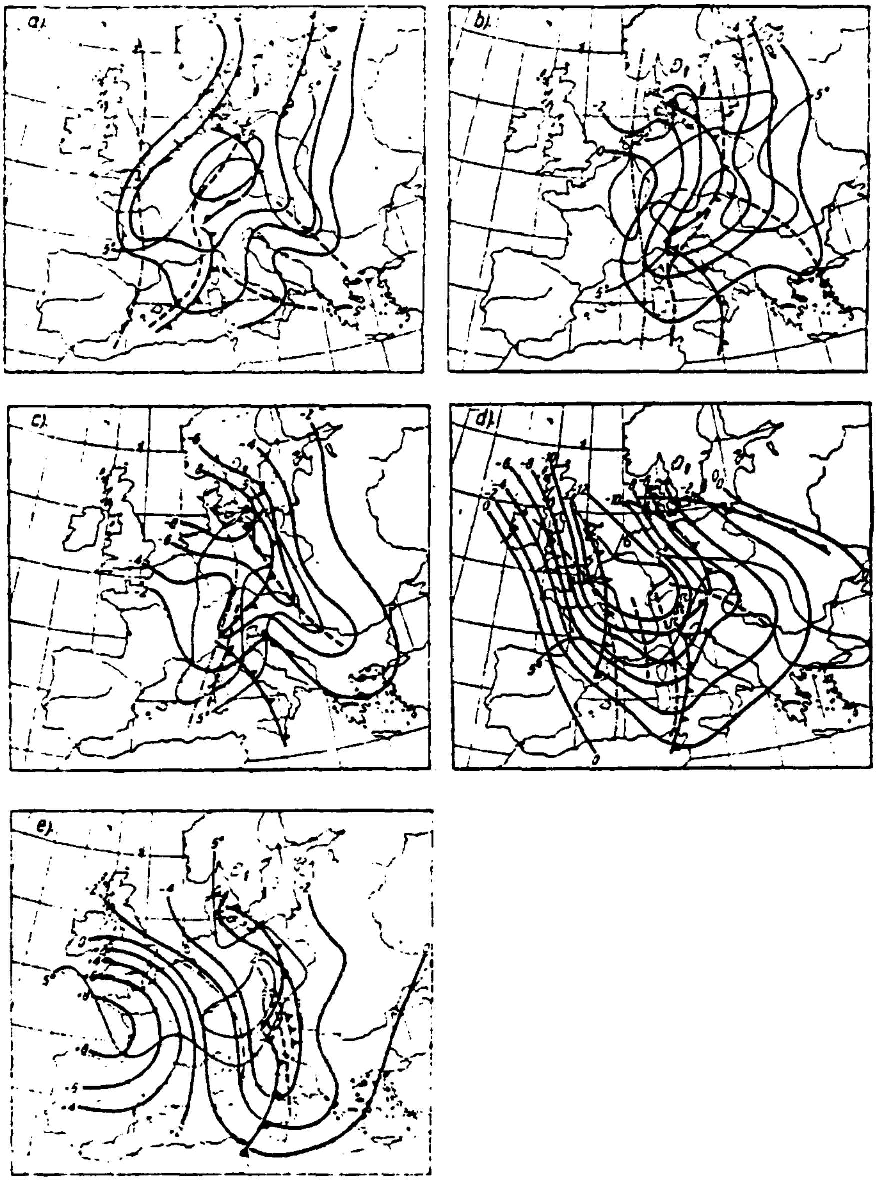


Fig. 74a) Distribution of the 24-hour isallohypses of the 850 mb surface on 12 August 1960, 00 hours GMT. The thick, solid lines are isallohypses, the thin, solid lines are curves of a dewpoint depression of 5°, the dashed lines are trough lines, the dot-dash line is the trough line of the 500 mb surface; b) Distribution of the 24-hour isallohypses of the 850 mb surface on 19 August 1960, 00 hours GMT; c) Distribution of the 24-hour isallohypses of 850 mb surface on 8 August 1963, 00 hours GMT; d) Distribution of the 24-hour isallohypses of the 850 mb surface on 13 July 1961, 00 hours GMT; e) Distribution of the 24-hour isallohypses of the 850 mb surface on 13 July 1961, 12 hours GMT (designations as in fig. 74a).

With regard to the timing of formation of the instability lines, we obtained the results presented in table 51 after application of the principles of George. From the table we see that a decrease in the spacing between the trough lines, gives a decline in the time difference between the timing of the topography chart and that of formation of the instability line. According to the charts, the intensification of the instability lines occurs within the Carpathian basin, but the areas of the Istrian peninsula and Slovenia played a role in their generation. The spatial shift of the lines can probably be attributed to the use of 24-hour isallohypses.

Table 51.

Zeitpunkt der Fälle 1	2 Mittlerer Abstand der Troglinien 850 und 500 mb (Km)	3 Entstehungszeit (Stunden)
12. Aug. 1960, 00 GMT (24-stündige 4 Isallohypsen)	1400	21
19. Aug. 1961, 00 GMT (24-stündige 4 Isallohypsen)	1100	9
8. Aug. 1961, 00 GMT (24-stündige 4 Isallohypsen)	700	4
13. July 1961, 00 GMT (24-stündige 4 Isallohypsen)	1200	12
13. July 1961, 12 GMT (12-stündige 4 Isallohypsen)	340	5 schon ausgebildet

Key: 1-timing of the cases 2-average spacing of trough lines 850 and 500 mb (km) 3-generation time (hours) 4-24-hour isallohypses 5-already formed

The case of 13 July 1961 is very instructive. In fig. 74d, a 24-hour isallohypsis chart is visible which has the trough line of the isallohypsis 850 mb inside the moist tongue above the Carpathian basin. In fig. 74e, the 12-hour isallohypsis field at 12 hours GMT is presented for the day in question. At this time, the instability line had just formed, the axis of the trough and the closed isallohypsis center lie near the line's formation point. By comparing the two charts, one draws the conclusion that a more accurate picture of the point of generation of the instability lines could be obtained if 12-hour isallohypsis fields were used. Due to the obtained results, it can be stated that in the presented cases, the conditions are favorable for the formation of an instability line in the Carpathian region.

From the obtained results we will not draw any further conclusions, but it is certain that the discussed method offers a simplification in the diagnosis of the synoptic situation.

In connection with the evaluation of this method, it must be mentioned that George does not discuss the theoretical basis of his method. But it can be presumed that the convergence regions of the friction layer are indicated by the trough lines of the isohypses of the 850 mb surface. Now the thunderstorm-triggering effect of these regions is well-known.

In summary, we find that a prediction of instability lines is possible today only by a synoptic, empirical method. Above all, near-ground and in the lower troposphere, the presence of a macrosynoptic situation favorable for the formation of a line must be found. With regard to the fact that the geographic location of the generation is known from the previous investigations, by means of a thorough analysis of this region performed on the basis of the mesosynoptic considerations, the line can be detected very early in the course of its development, provided that synoptic information of this region is available in short time intervals. Now the forming line can be subjected to a translation with the average speed computed from the material of the last 10 years (60-70 km/hour), and on this basis, a timely warning can be issued for the Lake Balaton region. But it is of greatest importance that the weather situations favorable to the formation of an instability line be subjected to another examination to permit needed experience to be gained for the preparation of objective, synoptic prediction methods.

7. Structure of Air and Water Temperatures in the Area of Lake Balaton

7.1 Introduction (G. Götz)

In this section, some studies are reported which pertain to the temperature conditions of lake water and the bank region, and to the interaction of water and air temperatures. It should be stated in advance that these investigations were initiated primarily by needs which arose in the course of operation of the Storm Warning Service, and conseqeuntly these investigations do not use the methods and requirements usually applied to the study of mesoclimatic objects.

Our intention in implementing these studies was to provide some help to synopticians of the Storm Warning Service for those cases where the problem is to give a short-term prediction (with a validity of 12 hours) or to report corresponding information.

7.2 Water Temperatures of Lake Balaton (B. Böjti)

Concerning the water temperature conditions of Lake Balaton--which is one of the largest inland lakes of Central Europe having a surface area of nearly 600 km^2--we have today only insufficient data. The goal of this work was to offer an overview of the lake's water temperature conditions on the basis of existing data and to describe the annual and daily temperature profile.

As background to the calculations, we used the following sources: Water temperature measurements at the Tihany Climatological Station of the Central Meteorological Institute from the years 1929-1958; water temperature measurements in Siófok 1946-1954, and the hourly water temperature measurements in Siófok taken in the summer of 1962 within the frame of the Balaton research.

The water temperature observations must be handled with all due caution because of the inaccuracy of the instruments and methods, and because of errors caused by differences in the measuring schedules, and lastly, because of individual errors of the observer. The determination of the water temperature uses an immersion method. The temperature of the water sample--taken from great depth in an open or closed thermobottle, can undergo a certain change up to the moment it is read. The smaller the extracted amount of water and the smaller the inertia of the thermometer used to measure its temperature, the greater can the errors be. Reliable, accurate water temperature data can only be obtained by using an electric thermometer which is always placed at that point whose temperature is to be measured.

A comparison of water temperatures with air temperatures likewise encounters great difficulties, since the observation schedules in most cases do not agree with the climatological observation. After an overview of the Winter water temperature measurements, one becomes more convinced that the evaluation of available data requires special care. Among the water temperatures from the winter, we find repeat-

edly values of -4 or -5 ^{o}C, which naturally cannot be seen as real values, since Lake Balaton is a freshwater lake and cannot have such temperatures.

For the shallow lakes of the temperate belt it is common that in summer and in daytime, the water surface is warmer than the lower layers, whereas in winter, the opposite is true. The density of the water reaches its maximum at +4 ^{o}C. Now if the water mass cools off in Autumn, then these masses sink into the depths, according to their specific weight and their place is taken by warmer water which rises to the surface through convection. The homothermal state occurs in the lakes at a temperature of +4^{o}. Further cooling is no longer a consequence of convective and dynamic motion, but it is caused rather by thermal conduction. Upon surface cooling below 0^{o}, ice begins to form. These statements also relate to Lake Balaton, although, due to the low water depth, the water is mixed very quickly due to wave motion, and consequently the above statements loose their validity in the case of strong winds.

The earliest ice-day for Lake Balaton is December 8, and the last occurs on March 30. In the meantime, i.e. in the winter, the water temperatures remain lower than the critical homothermal temperature value. The water temperature data coming from this period can only be used for information.

Accordingly, in the lake water a temperature layering can be found like that in an air column. When moving into the depths, in summer one finds colder, and in winter, warmer layers. This temperature layering can also be found in the shallow Lake Balaton. For a sharper picture of the temperature layering, the deeper water is suitable. Only measurements performed at the deep point of the lake, the "Well of Tihany", gave gradient values of 0.6 to 1.8^{o} between the surface and the lake bed. We consider it probable that on bright, calm summer days, even greater gradients can occur, but this can only be proven when a proper set of measurements has been taken. The following factors are very important for the structure of water temperature: Radiation, air temperature, wind, vaporization, precipitation.

The annual profile of air temperature in Siófok and Tihany and of the water temperature from the measurements in Tihany were presented in table 52. In the table there are monthly and annual averages computed from data over a 30-year period. As we see, it is characteristic for the annual profile of temperature, that from the time the ice melts, a stepwise warming of the water occurs. The warmest month is July, then there is a temperature drop until December.

If the random measurements are also taken into account, then the previous, max. water temperature on the open lake was 31^{o}. If the lowest temperature is assumed to be 0^{o}, then this means that the absolute amplitude of the annual profile of water temperature in Lake Balaton has the value of 31^{o}.

Table 52. Monthly Average of Air and Water Temperatures (1929-1958)

1 Monat	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	Jahr 2
3 Lufttemperatur in Siófok	—0,9	0,7	5,8	10,9	16,3	19,4	21,4	20,4	16,4	11,0	5,2	1,1	10,6
4 Lufttemperatur in Tihany	—1,0	0,8	5,6	11,0	16,3	19,6	21,8	20,8	17,2	11,7	5,5	1,4	10,9
5 Wassertemperatur in Tihany	0,0	0,0	4,5	11,5	17,6	21,9	23,2	22,9	19,2	13,3	6,9	2,0	11,9

Key: 1-month 2-year 3-air temp. in Siófok 4-air temp. in Tihany
5-water temp. in Tihany

The annual average of water temperature in Lake Balaton is higher than the average air temperature at the two bank stations: The extra amount is 1^{o} for Tihany and 1.5^{o} for Siófok. The water of the lake has a higher temperature than the air on average from 15 April to 15 Dec. If we note that the temperature data do not come from daily averages, but from observations at 7 a.m., i.e. at the time when the water temperature has its lowest value, then this finding is acceptable. A closer observation of water temperatures in January and February is out of the question, since during this time the lake is covered with ice. In March and April, the air is even warmer, on average, than the water. The turning point is around April 15. This can be explained since in early Spring, large quantities of heat are needed to melt the ice, and since

in early Spring, the temperature of the inland and the air above it rises at a much greater rate than the temperature of the water mass which has a high heat capacity.

In these individual cases, the influence of the lake on the temperatures of the bank region was not determined by a positive, extra amount for the water temperature, but by the difference of air temperatures present between the air over the water and that over solid ground. In a summarization of the snyoptic messages reported hourly at the Siófok observatory, it is frequently observed that the air temperatures are subject to a powerful change within a short time, and these important temperature jumps are always linked with a wind rotation. Such cases occur mainly in the midday hours with their greatest sharpness, at a time when the air over the land has a much stronger warming than over the water. If there is a wind rotation at this time which is directed toward the bank, then a temperature jump of more than 5 degrees can occur. The wind on 24 March 1961 turned from the direction NNW to ESE between 13 and 14 hours, and there was a temperature increase from 7.5° to 13.8°. The water temperature at this time was 6.8°.

For a shallow lake like Lake Balaton, it is characteristic that the water temperature reacts very quickly to a larger and lasting change in air temperature: The temperatures of the water of Lake Balaton follow with great certainty the changes in air temperature. For proof of this fact, a correlation calculation was performed between the water and air temperature data of the stations at Keszthely, Siófok and Tihany. For these calculations, we have added the observations performed in the course of 1961 at 07 hours. As an example of our results, the ranking correlation results for two variables are reported for Siófok:

$$r = 1 - \frac{6 \Sigma d^2}{n(n^2 - 1)} = 1 - \frac{6\,(2689.89)}{308\,(94864 - 1)} = 1 - 0.007 = 0{,}993$$

whereby d^2 is the square of the temperature difference between the air and water temperatures measured at the same time, n is the number of cases and r is the ranking factor of the correlation. The specific character of this relation can be found immediately if the

standard criteria for evaluation of a correlation factor are applied.

In the interest of further proof, those values of water and air temperature were chosen in cases of a landward wind in Siófok. From these 135 cases, a value of the correlation factor of 0.984 was computed. A control calculation was performed by means of the formula:

$$\varrho_r = \frac{6\sqrt{\left(d^4 - \frac{1}{n} d^2\right)^2}}{n\,(n^2 - 1)}$$

whereby ϱ_r is the probable error of the correlation factor. Since the value of the correlation factor is greater than 6 times the probable error, our calculations can be considered acceptable. Thus the close relation by which the air and water temperatures are linked is proven. But it must be mentioned that the months of early Spring and late Autumn are an exception, since in this period the change in air temperature takes place with such speed that the water temperatures cannot completely follow, in spite of the small water mass. This circumstance is why in the Spring months, a cooling effect on the bank landings by the lake is often noted, whereas in the Autumn months, a warming effect is frequently found, and this effect is sometimes expressed in the average values.

The statements above are invalid when a short-term, but very quick change in air temperatures takes place which the water temperature cannot follow. As an example, the cold front of 15 July 1962 had an air temperature at 16 hours to 21 hours of 7 degrees, but the air temperature [sic] decreased only by 0.2 degrees. As a second example, the cold front of 16 August 1962 had a 13 degree decrease in air temperature over 12 hours, but the decrease in water temperature was only 3 degrees.

The determination of the daily profile of water temperature of Lake Balaton is a rather difficult task due to the lack of sufficient data. Some help is provided by the hourly observations of the Siófok observatory taken in 1962. On this basis, the daily profile of water temperature is shown in fig. 75. From 19 hours evening to 11 hours morning, the water temperature is higher than the air temperature

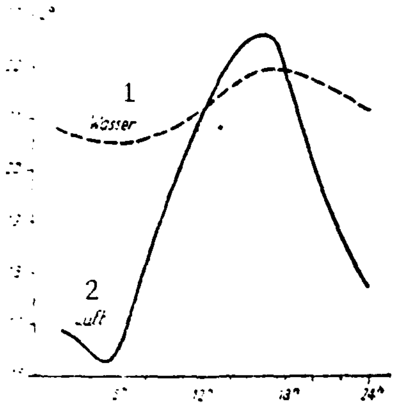


Fig. 75. Hourly Structure of Air and Water Temperatures in Siófok, July 1962

Key: 1-water   2-air

whereas during the time from 11 hours to 19 hours, a higher air temperature is characteristic. In the daily course of water temperature, two sections can be distinguished: An ascending branch from 7 to 19 hours, and a descending one from 19 hours to 07 hours. The average daily amplitude is 4 degrees. Concerning the other details of these measurements, the reader is referred to section 7.4.

These paragraphs do not answer all questions about water temperatures of Lake Balaton. The relations between precipitation, incoming and outgoing radiation and the air and water temperatures are a whole series of other problems. But these factors can only be investigated on the basis of permanent and timely temperature measurements.

7.3 Daily Course of Temperature at the Bank of Lake Balaton (T. Tänczer)

Knowledge of the daily temperature profile at a given location is important for synoptics and climatology. If we form averages of many years, then the daily temperature profile shows the effect of solar radiation in the foreground, and leads to the formation of a curve resembling a sine-shape. But if individual days are considered, then the picture is by no means so regular. Other effects (including primarily cloud effects and temperature advection) can lead to a significant change of this picture, even though a daily, positive half wave and a nightly negative half wave can be distinguished in most cases. The two factors mentioned above vary from case to case so that practically for every day, we get a different type of daily profile curve. But this finding is not valid for days when there is no significant temperature advection and few clouds. On such days,

the incoming and outgoing radiation are unimpeded and the daily course of temperature can be viewed as an exclusive function of the season and geographic conditions of the particular location.

In the present discussion, we will attempt to determine the daily profile of temperature in the summer period and on front-free, bright days on the basis of temperature observations at the Siófok observatory. The investigation was based on observations from 5 years (1960-1964) for the period of 15 May-31 July. In selecting this period of time, we were guided by the fact that we wanted a period which was symmetric around the summer solstice and which had no significant changes in solar attitude and length of day. Now in the named period, the noon solar attitude in Siófok exceeds the value of 60° and the length of day is more than 14 hours, so there are no essential differences in the incoming and outgoing radiation. During this time, those days were selected when the amount of total cloudiness did not exceed the value of 3 Okta and there was no passage of a front. In the course of 5 years, we succeeded in finding 37 such days.

On the basis of the hourly temperature changes, we determined the average daily profile for these days. The temperature changes occurring in the night between 22-01 hours and between 01-04 hours, were combined since hours in between sometimes had no data and for these time segments, a greater resolution would have little practical benefit.

The computed changes from hour to hour are reported in table 53.

Table 53.

1 Stunde	01—04	04—05	05—06	06—07	07—08	08—09	09—10	10—11	11—12	12—13
2 Grade	—1,7	—0,3	+1,0	+2,3	+1,5	+1,3	+0,9	+1,3	+0,9	+0,9
1 Stunde	13—14	14—15	15—16	16—17	17—18	18—19	19—20	20—21	21—22	22—01
2 Grade	+0,6	+0,5	+0,3	+0,0	—0,4	—1,7	—2,1	—1,2	—1,0	—2,4

Key: 1-hour 2-degrees

For the sake of simplicity, the obtained temperature profile is also illustrated graphically (fig. 76). Above all, it is striking that the daily profile of temperature shows a deviation compared

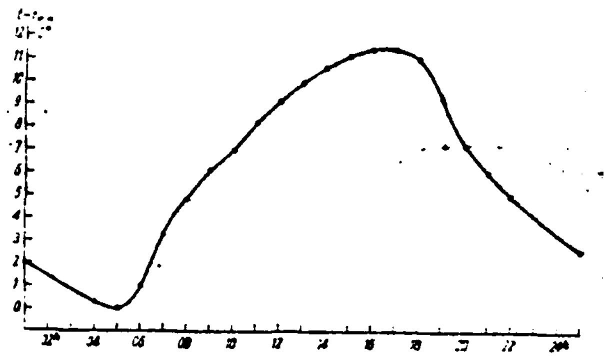


Fig. 76. Daily Profile of Temperature in Siófok on Front-free, Bright Days (with cloudiness $N \leq 3/10$), for the Period of 15 May-31 July in the Years 1960-1964.

to the sine wave. This deviation is due to the rapid increase or decrease in temperature occurring on the sudden onset of radiation and its equally sudden disappearance. At this time, the warming or cooling extends only to the air quantity located directly over the ground surface. Through the onset of convection, the warming process is extended to a larger air layer. Accordingly, in the course of the forenoon hours, there is a more significant slowing of warming. According to the figure, the first stagnation of the warming process occurs at about 9 hours, and it could be a result of the beginning of cluster-cloud formation. Since on the studied days there was only a little cluster cloud formation (up to 8 Okta), this stagnation in warming is not as pronounced as in the preceeding section by Endrödi and Götz. The curves show the effects exerted by Lake Balaton. This influence is most glaring in the shift of the timing of the temperature maximum. According to the curve, the temperature maximum in Siófok occurs in the time between 16 and 17 hours, in contrast to other stations, where the time is between 14 and 15 hours. This lengthening of the warming period must be attributed to the moderating effect of the lake, and this appears in a prominent manner on sunny days. Under such conditions, the forming, local air circulation near the ground causes landward winds by day, and offshore winds by night. The alternation of air movements causes the daily amplitude of temperature to have a relatively high value, in our case, of 12.3°, in spite of the well-known moderating effect of the lake. In the descending branch of the curve, at 20 hours we have a stagnation which is probably related with the onset of dew

formation. If the starting point of the curve is compared with the end point, then we find that the temperature at the endpoint is greater by 0.6^{o} than the starting point. This means that within the studied time, the air temperature is raised on average by 0.6^{o} daily by the radiation of near-ground air layers. However, if this value is greater, then in anticyclonal weather situations, the appearance of such atmospheric conditions must be expected.

As we see from the number of studied cases, it is fairly unusual for these assumptions to be fully met. Nevertheless, this calculation does have a practical value. The daily profile described here represents so to speak, the most extreme possibilities which are reduced by the cloud cover in accord with the cloud quantity. Here, the effects of a temperature advection can be estimated exclusively on the basis of the synoptic chart.

7.4 Interaction of Water and Air Temperature on Sunny Days in the Area of Lake Balaton (G. Endrödi and G. Götz)

Between the ground and air air flowing over it, there is a constant interaction. These influences try to reduce the differences caused by static and kinematic quantities. Due to the solid ground, air movements over it are braked by friction, the wind causes waves on the water and there is an uninterrupted exchange of heat and moisture between the lower air layers and the surface of the ground. The investigation of this process is an important branch of meteorological research, since without a detailed knowledge of it, neither the large-area questions (problems of energy balance of the atmosphere, or the properties of general air circulation), nor the data of quite small-area, micrometeorologic observations can be interpreted correctly.

The process of this interaction is a very complex, interrelated group of phenomena--except for some trivial cases. It is dependent on the duration of the influencing factors, on the differences initially present in the values of the characteristic quantities, and above all, on differences existing with regard to the spatial size of the two systems present on both sides of the interface, and on differences relating to the available energy reserves. Through

these factors, the results of the influences emanating from both sides can in general, be determined, so that e.g. a snow cover always has a cooling effect on an air mass coming from the direction of the lake, and the practically inexhaustable water mass of the oceans lead under all circumstances to an increase in the moisture content of the particular air mass, or the orography leads in all cases to a weakening of the wind via friction. Due to the enumerated influences resulting from the extent and duration of them, the properties of an air mass can be affected, and it follows that they must be included in the framework of synoptic meteorology.

A somewhat more apparent, but no less important question for studies on this scale, is the interaction between ground and air, for example for the Lake Balaton region, where the temporal and spatial dimensions, and the exchangeable energy supplies are significantly smaller. The influences exerted by the lake on its environment have been examined by several authors, but the synoptic side of this question is still important. A prediction of the daily extreme air temperature at the bank, and a prediction of the water temperature are recurring tasks over the summer, which are often problemmatic for the synopticians of the Lake Balaton Storm Warning Service. The problem can be solved exclusively on the basis of a precise knowledge of the interactions existing between the water and air temperature.

For the magnitude of conditions existing in this case, the interaction in both directions can be equally great, and for this reason, the question must be a two-fold one: How is the lake water affected by the air above it, and what kind of influence is exerted by the water surface at a given time, on the air moving over it? The treatment of these two questions requires a differing method. The influence by the air temperature on the structure of the water temperature, can be investigated by means of an observation series of air and water temperatures, based on measurements performed at the particular site. The basic viewpoint of this study consists in the question of how a sudden cooling or a lasting warming of the air follows the changes in water temperature, and how the daily profile of air and water temperatures are structured on bright and calm days. The second, opposing question is answered such that

the thermograph recordings at two opposite-bank stations were compared, whose connecting line coincides with the wind direction, and the dependence of this thermogram on the water-air temperature difference and on the wind speed (which gives a measure for the time of the air over the water) can be studied. This duality of the problem naturally does not mean that we are dealing with a one-sided effect in either case. In both cases there is an interaction, the duality is justified solely in the fact that differences occur in the magnitudes of the two, opposing effects.

In the present paper, only one of the presented viewpoints will be discussed: How are the daily profiles of water temperature at various depths, and the air temperature at the bank of the lake and in a certain distance inland, structured? Within the frame of the research program for Lake Balaton, which was determined by our climatology department, in Siófok from 5 July to 31 Oct. of 1962, the following measurements were performed hourly: Water temperature about 15 m from the bank, at depths 0, 50, 100 and 160 cm (on the lake bed); air temperatures at 50 and 200 cm above the water and (until 31 August) at about 1.2 km inland from the bank, also at heights of 50 and 200 cm. The results, supplemented by the hourly observations of the Siófok meteorologic station--performed in a regulation thermometer hut at 10 m from the bank--permit a thorough analysis of the daily profile. The starting values were selected in the individual months from 15 such days, on which a sunny weather situation prevailed and the conditions were not greatly affected by clouds, precipitation, winds and frontal passage. A model of the daily profile of temperature for four different months is presented in fig. 77.

If any of the studied months is taken into consideration, it is immediately striking that--as expected--the smallest values of the temperature amplitude occur in the water, and the greatest values are inland or at the bank stations (table 54). If we examine the chronology of the temperature amplitude, we find that the greatest amplitude of air temperature occurs in October, whereas for the water, it occurs in August and attains its minimum amplitudes in October. Consequently, in October we have the greatest temperature

contrasts between air and water, since at this time an amplitude difference of 8.8° exists between water surface and 200 cm layer, whereas in July, the same quantity has a value of only 5.2°.

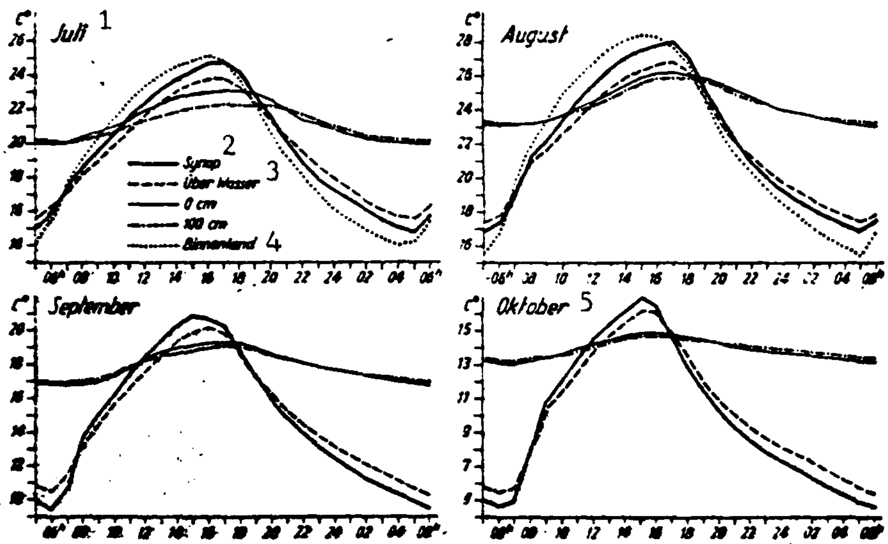


Fig. 77. Daily Profile of Air and Water Temperature on Sunny Days in Siófok

Key; 1-July 2-synoptic 3-over water 4-inland 5-October

Table 54. Daily Amplitude of Temperature in the Lake Balaton Area

Meßort 1	2 Juli	Aug.	Sept.	Okt. 3
4 Wasser 100 cm	2,2	2,8	2,2	1,5
4 Wasser 0 cm	3,1	3,1	2,5	1,8
5 Über Wasser 200 cm	8,3	9,3	9,6	10,6
6 Uferstation	9,8	11,0	11,4	12,2
7 Binnenlandstation	11,1	12,9	—	—

Key: 1-measurement location 2-July 3-Oct. 4-water 5-over water, 200 cm 6-bank station 7-inland station

The explanation of this fact must lie in the different heat balance of these two media. In the course of the lengthening Autumn nights, a greater cooling occurs in the air than in the water, which stores heat from the summer months, and in the shorter and low-radiation days, it is less able to heat the air than in summer. Conversely, the relatively small water mass of Lake Balaton does not supply enough heat to replace the heat lost by out-radiation from air at the 200 cm level; to do so, the amplitude would have to be no

ian region.

greater than on sunny days in the summer. It is an interesting fact that the daily temperature amplitude over the water in all four months is nearly the same amount less ($1.6-1.8^{o}$) than at the bank stations. This fact must be attributed to the moderating effect of the water, and based on the number data, one obtains the impression that the extent of this influence exerted by the water on the air above, has the same amount for each month.

If the annual profiles of temperature based on averages for the selected days are examined, then we find that from July to October, the timespan in which the water is colder than the air at the bank, is reduced stepwise. Whereas in July, this time extends from 6-7 hours to 20-21 hours, the two temperature curves intersect in October at the times 8 hours and 16-17 hours. Thus, a reduction of this time span is caused primarily since the time of arrival of max. values in the summer months of 16-17 hours, is shifted in Autumn to 15 hours, the cooling begins over the water at an earlier time and consequently, the change in sign of the temperature difference between the two air temperatures occurs at an earlier time. From the above discussion, it follows that the temperature difference between the air temperatures over the water and over the bank assume their maximum at the time when the extreme values occur: At the time of the minimum, it is $0.5-1^{o}$, and at the time of the maximum, $0.8-1.2^{o}$. On average, for 60 cases, in the air over the water, we have a reduction of extreme values of temperature of about one degree due to the moderating influence of the lake.

If the data of land stations is included, then we find that both the warming, and the cooling occur at an earlier time and the maximum occurs one to two hours earlier than on the bank or over the water--as expected. In the course of the two summer months with observed data from this station, we see that the temperature curve of this station differs sharply from the curves of the other stations. With increasing distance from Lake Balaton, there is an increasing growth in the inland influence: Both the cooling as well as the warming, is greater. The temperature differences between the inland station and the bank station, both between the air over the

by needs which arose in the course of operation of the Storm Warning Service, and conseqeuntly these investigations do not use the methods and requirements usually applied to the study of mesoclimatic objects.

inland and the air over the water, attain their greatest values at the time of intensive warming and in the course of cooling before sunup, and in August, in the times 9-15 hours and 3-5 hours, the latter temperature difference can exceed a value of 2°.

From the structure of the water temperature it can be found that this temperature has a very uniform daily profile. The temperature difference between water and air is greatest in the time of cooling, and as the days get shorter, there is a constant increase in this difference. In October, at the time of the day minima, the water is 7-8 degrees warmer than its environ, whereas in the early afternoon hours, is stays only 1 to 2 degrees colder than the air.

The differences resulting between the intensities of warming and cooling can be described by numeric data. If the temperature amplitude is divided by the time passing between the two settings of the thermometer, then we obtain the speeds of warming (+) and cooling (-) in units of degrees/hour, as illustrated in table 55.

Table 55. Speed of Warming and Cooling (Degrees/Hour) in the Balaton Area

Messort 1	2 Juli		August		September		Oktober 3	
	+	—	+	—	+	—	+	—
4 Wasser 100 cm	0,22	—0,20	0,31	—0,23	0,24	—0,17	0,20	—0,11
4 Wasser 0 cm	0,28	—0,23	0,31	—0,24	0,28	—0,20	0,22	—0,13
5 Über Wasser 200 cm	0,74	—0,64	0,78	—0,78	0,96	—0,69	1,18	—0,71
6 Uferstation	0,90	—0,32	0,92	—0,92	1,27	—0,76	1,35	—0,81
7 Binnenlandstation	0,92	—0,92	1,29	—0,92	—	—	—	—

Key: 1-measurement location 2-July 3-October 4-water 5-over water 6-bank station 7-inland station

Naturally, the speeds of warming and cooling in all months have their lowest values for the water; for the air, the speed of temperature changes is about three times faster. If we compare the observations of the bank station with those over water, we always find values over the lake which are smaller by several tenths of a degree, which must doubtless be attributed to the effect of the water on the air. This influence is greatest in the air layer lying next to the water, but it can also be detected at the bank.

In a more detailed investigation of the ascending branch of the daily profile, two other effects can be found in the air, and consequently in the water to a certain extent. These are not related to the particular set of problems here, but are worth mentioning. The one circumstance is that in the time between 9-10 hours--especially in summer--a certain retreat in warming intensity is found which can be explained by the convection beginning at this time. The heat losses of the near-ground air layer caused by this, are soon compensated by the increasing insulation. The second notable fact is the exceptionally strong intensity of warming occurring in September between 6 and 8 hours and in October between 7 and 9 hours. Of course, in all months, we find that in the first hour or first two hours after arrival of the minimum, the strongest warming occurs, but this warming is enhanced in the autumn radiation period since the thin, near-ground inversion layers form with increasing frequency and are then quickly dissolved by the solar radiation.

Table 56. Some Special Characteristics of the Temperature Relations in the Balaton Region.

		1 Juli	August	Sept.	Okt. 2
3 Zeitpunkt des Eintritts von T_{max}	4 Uferstation	16—17^h	17^h	15—16^h	15^h
	5 Wasser 100 cm	17—19^h	16—18^h	16—17^h	15—16^h
3 Zeitpunkt des Eintritts von T_{min}	4 Uferstation	04—05^h	05^h	06^h	06—07^h
	5 Wasser 100 cm	05—06^h	06—07^h	06—07^h	06—07^h
$T_{max} = T_{07^h} +$	4 Uferstation	+7,3°	+8,6°	+10,1°	+12,0°
	5 Wasser 100 cm	+2,2°	+2,8°	+ 2,2°	+ 1,5°
$T_{min} = T_{19^h} +$	4 Uferstation	—7,9°	—8,2°	—7,8°	—6,9°
	5 Wasser 100 cm	—2,2°	—2,6°	—1,9°	—1,2°
6 Wassertemperatur in 100 cm Tiefe	< Ufertemperatur 7	10—20^h	10—19^h	11—18^h	11—17^h

Key: 1-July 2-Oct. 3-time of arrival of 4-bank station 5-water 6-water temperature at 100 cm depth 7-bank temperature

In the present study, we did not endeavor to cover all details of the interactions of air and water. The investigation was aimed primarily at questions concerning everyday problems in the work of the Siófok Observatory. We intended to help the synopticians by the data in table 56. Naturally, these values, as all other results of this study, are affected by the peculiarities of the weather picture in the observation period and thus must never be forgotten in the evaluation of the results. Thus, compared to our present

AD-A114 322 FOREIGN TECHNOLOGY DIV WRIGHT-PATTERSON AFB OH F/G 4/2
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picture, differences can occur which relate not only to a weather picture different from the radiation weather picture (one-sided warming or cooling due to the arrival of clouds, precipitation and wind rotation), but also certain deviations can be perceived when, at a later time, a more representative investigation of these phenomena is undertaken over a longer observation period.

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